Archived version from NCDOCKS Institutional Repository http://libres.uncg.edu/ir/asu/

Appalachian STATE UNIVERSITY
BOONE, NORTH CAROLINA

Southeastern Geology: Volume 30, No. 3
December 1989

Edited by: S. Duncan Heron, Jr.

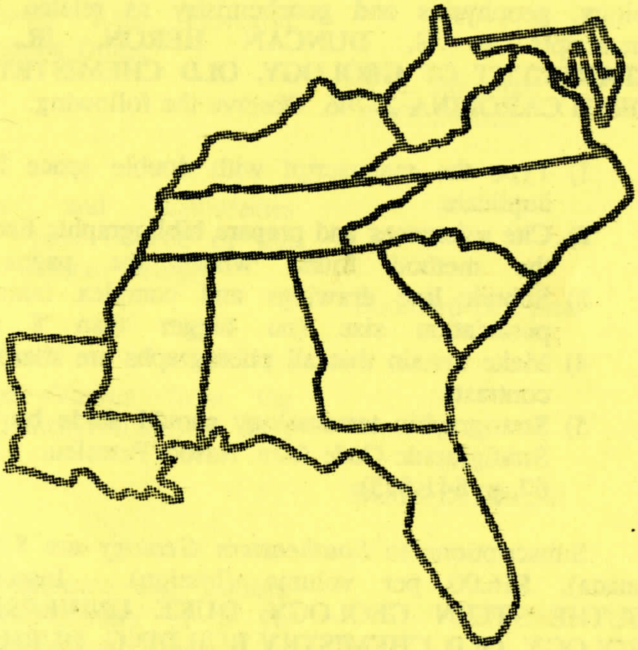
Abstract

Academic journal published quarterly by the Department of Geology, Duke University.

Heron, Jr., S. (1989). Southeastern Geology, Vol. 30 No. 3, December 1989. Permission to re-print granted by Duncan Heron via Steve Hageman, Professor of Geology, Dept. of Geological & Environmental Sciences, Appalachian State University.

SERIALS DEPARTMENT
APPALACHIAN STATE UNIV. LIBRARY
BOONE NC

SOUTHEASTERN GEOLOGY



PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

VOL. 30, NO. 3 DECEMBER 1989

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY

Editor in Chief:
S. Duncan Heron, Jr.

Managing Editor:
James W. Clarke

This journal publishes the results of original research on all phases of geology, geophysics and geochemistry as related to the Southeast. Send manuscripts to **S. DUNCAN HERON, JR., DUKE UNIVERSITY, DEPARTMENT OF GEOLOGY, OLD CHEMISTRY BUILDING, DURHAM, NORTH CAROLINA 27706.** Observe the following:

1) Type the manuscript with double space lines and submit in duplicate.
2) Cite references and prepare bibliographic lists in accordance with the method found within the pages of this journal.
3) Submit line drawings and complex tables reduced to final publication size (no bigger than 8 x 5 1/8 inches).
4) Make certain that all photographs are sharp, clear, and of good contrast.
5) Stratigraphic terminology should abide by the North American Stratigraphic Code (Am. Assoc. Petroleum Geologists Bulletin, v. 67, p. 841-875).

Subscriptions to *Southeastern Geology* are $12.00 per volume (US and Canada), $16.00 per volume (foreign). Inquires should be sent to: **SOUTHEASTERN GEOLOGY, DUKE UNIVERSITY, DEPARTMENT OF GEOLOGY, OLD CHEMISTRY BUILDING, DURHAM, NORTH CAROLINA 27706.** Make checks payable to: *Southeastern Geology*.

SOUTHEASTERN GEOLOGY

Table of Contents

Vol. 30, No. 3

December 1989

THE NASHVILLE DOME - AN ISOSTATICALLY INDUCED EROSIONAL STRUCTURE - AND THE CUMBERLAND PLATEAU DOME - AN ISOSTATICALLY SUPPRESSED LATE PALEOZOIC EXTENSION OF THE JESSAMINE DOME

A. L. REESMAN
R. G. STEARNS

Department of Geology
Vanderbilt University
Nashville, TN 37235

ABSTRACT

Isostatically adjusted structure maps, or no-load surfaces, were constructed on the base of the Chattanooga Shale. These maps suggest that the structure of the Nashville dome can be interpreted as a by-product of isostatic adjustment to late Tertiary to Recent erosion and has essentially no "tectonic" (non-isostatic) component, whereas an isostatically suppressed "tectonic" dome is buried beneath the Cumberland Plateau. Although emphasized by isostatic adjustment to erosion, the Jessamine dome (unlike the Nashville dome) has a significant "tectonic" component. On a no-load map the Jessamine dome and the Cumberland Plateau dome connect to form a structural arch that postdates the deposition of the Chattanooga-New Albany Shale. Patterns of post-Chattanooga sedimentation suggest that this arch influenced Upper Mississippian and Pennsylvanian sedimentation.

The elongated southern portion of the Cumberland Plateau dome parallels Sequatchie Valley and nearby western thrusts of the Valley and Ridge province, but not the Pine Mountain block. The arch diverges northward to connect with the Jessamine dome. Coincidence between the Cumberland Plateau dome-Jessamine dome arch and the East Continent gravity and magnetic high suggests that the arch was controlled by Precambrian basement. Coincidence of its south end with the west limit of Appalachian thrusts suggests that it existed late enough into the Paleozoic to act as a buttress to thrusting.

INTRODUCTION

Central Tennessee and perhaps much of the Interior Low Plateau have been the site of thousands of feet of vertical movement, most of which occurred since deposition of the Upper Devonian Chattanooga Shale (Stearns and Reesman, 1986). Isostatic adjustment is certainly a factor in such vertical movement on a regional scale. Most structural closure on the Nashville dome also postdates the Chattanooga (Wilson and Stearns, 1963). This paper examines the structural effects in the development of basins and domes that can be defined using the present structure on the base of the Chattanooga Shale and its equivalents and tries to separate the isostatic from non-isostatic components in these structures. In particular, it explores the possibility that localized isostatic adjustments might account for the closure of the Nashville dome.

This study involves much of the Southcentral United States (Figure 1A) and, in particular, Middle Tennessee (Figure 1B) in the region of the Central Basin to the Cumberland Plateau. The Central Basin is a lowland area in Central Tennessee that is surrounded by the Highland Rim, a higher level of the Interior Low Plateaus (Figure 1B). The crest of the Nashville dome is precisely centered in the Central Basin with flanks of the Nashville dome extending beneath the Highland Rim. The flanks steepen away from the center of the basin near the escarpment that separates the basin from the rim. The correspondence between dome and basin makes the Central Basin a prime candidate to be a structurally-controlled basin much like elongated valleys in the Appalachians; however, the relationship between basin and dome can be interpreted another way; that is, the dome could be in large part an isostatic uplift responding to differential unloading by erosion.

At present, rates of unloading are markedly different between basin and rim. The basin is eroding about 3 times as fast as the surrounding rim, and the basin may have existed for only about 5 million years (Reesman and Godfrey, 1981). Maybe the dome is also about 5 million years old.

There are justifications for this study of isostatic adjustment. First, isostatic adjustment appears possible by analogy with the rebound of Lake Bonneville (Crittenden, 1963 a,b), an area of nearly similar size. Second, as we will show, it was successful in that calculated isostatic adjustments accounted for the structure of the Nashville dome. Third, it is useful, because isostatically adjusted surfaces resolved another geologically significant structure, the Cumberland Plateau dome, which connects to the Jessamine dome via the East Continent gravity and magnetic high.

The approach of this study is empirical. Land surface and Chattanooga structural elevations were averaged over various areas to calculate a variety of isostatically adjusted structure maps ("no-load" surfaces) on the base of the Chattanooga Shale. Isostatic adjustments involved removing load by backstripping and adding it by infilling to arrive at a calculated surface with a no-load condition on the unconformable pre-Chattanooga surface. Constructing no-load surfaces is a "layman's approach" to isostatic adjustment. This study was limited in scope and sophistication. All calculations were made by hand because the initial study was small and it grew slowly enough that a computer was never used.

The main focus of this study is on Tennessee, because our original goal was to try to estimate how the erosional growth of the Central Basin might influence the structure of the Nashville dome through isostatic rebound. The study area was enlarged to include Kentucky and northern Alabama in order to show how these adjustments might affect the Chattanooga surface on structures that are adjacent to the Nashville dome, e.g., the Jessamine dome and nearby parts of the Illinois, Appalachian, and Warrior basins. As a result, the illustrations in this study contain maps both of Tennessee and of the Southcentral United States, e.g., the physiographic maps of Figures 1A and B.

THE NASHVILLE DOME AND THE CENTRAL BASIN

Outliers of Chattanooga Shale occur across the Central Basin and provide control for this, the youngest mappable datum on the Nashville dome. The base of the formation was used because it is a regional unconformity and it was likely to have been isostatically adjusted in Late Devonian time. Figure 1 shows the

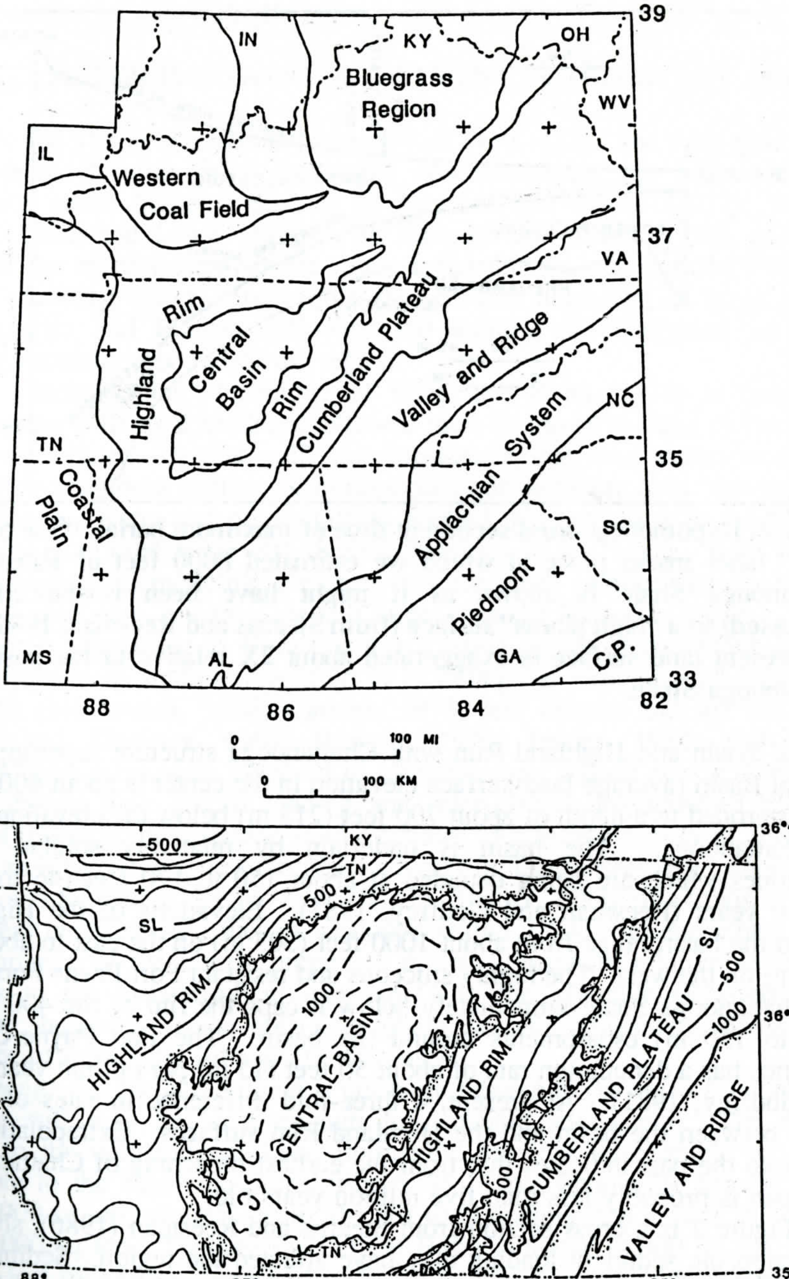


Figure 1. Physiographic provinces of the South Central United States (Figure 1A) and Central Tennessee (Figure 1B). Maps of these regions will reappear in many other figures. The structure on the base of the Chattanooga Shale is also shown in Figure 1B to depict the structure of the Nashville dome. Contours are dashed over the Central Basin where erosion has removed most of the shale. Notice how well the Nashville dome is centered over the Central Basin.

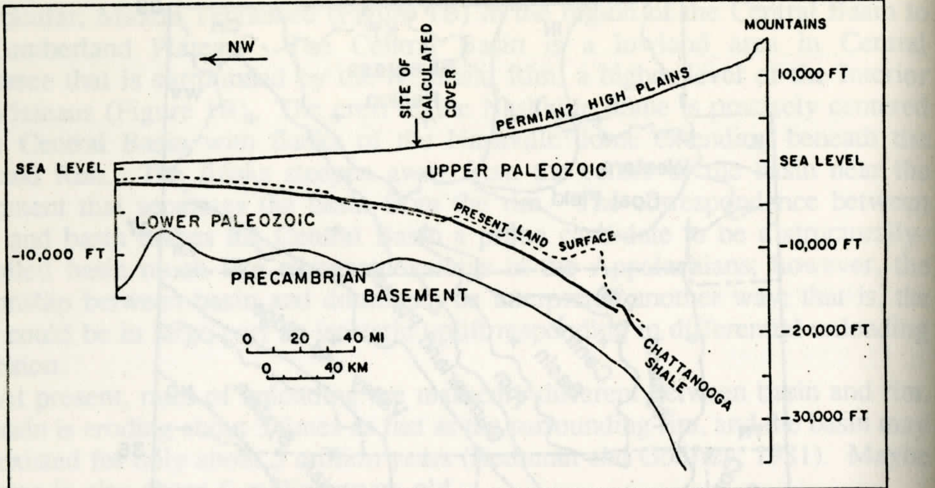


Figure 2. Hypothetical cross section at time of maximum burial. "Site of calculated cover" label marks place at which we estimated 8000 feet of Paleozoic cover. Chattanooga Shale is shown as it might have been isostatically adjusted (depressed) to a "high plains" surface (from Stearns and Reesman, 1986, Figure 8). The present land surface is exaggerated about 2X relative to the position of the Chattanooga Shale.

Central Basin and Highland Rim with Chattanooga structure superimposed. The Central Basin (average land-surface elevation in the center is about 600 feet or 184 m) has eroded to a depth of about 700 feet (213 m) below the elevation of outliers of Chattanooga. The basin is underlain by relatively soluble Ordovician limestones which are being denuded at about 150 to 170 feet (46 to 52 m) per million years (Reesman and Godfrey, 1981). Elevations of the Highland Rim around the basin range from about 1000 feet (305 m) on the east to about 850 feet (260 m) on the west. The highly siliceous and resistant Fort Payne Formation with the Chattanooga Shale immediately below it caps the rim of the 400 to 500 foot (120 to 150 m) escarpments around the basin. The Fort Payne-Chattanooga sequence has a denudation rate of about 50 feet (15 m) per million years (Reesman and Godfrey, 1981). At present, a three-fold difference in rates of denudation exists between the basin and the Highland Rim surfaces. Extrapolation of these rates into the past indicates that from the earliest breaching of Chattanooga Shale the basin is probably less than five million years old.

Figure 2 is a cross section from Stearns and Reesman (1986), showing how late Paleozoic sediment loading may have affected the region. Sediment loading was estimated to be about 7,000-8,000 feet (2100-2400 m) based upon temperature measurements on fluid inclusions from Lower and Middle Ordovician ore and mineral deposits in Middle Tennessee (Roedder, 1971), and from the rank of the coals on the Cumberland Plateau. We assumed that "high plains" type sediment, shed from the Appalachians, covered the region over the present Nashville dome during this period of time. Christopher Beaumont (written communication, 1986), using more refined estimates from the rank of the coals, suggested 8,000 feet (2,400 m). This regional loading and later unloading must have involved isostatic adjustment. Our estimates of the amount of vertical motion and the process of constructing Figure 2, lead to our attempt to evaluate the isostatic component in the present structure of the Nashville dome.

A CASE FOR ISOSTATIC ADJUSTMENT IN THIS STUDY AREA

Since the 1850's the isostatic concepts of Pratt (1855) and Airy (1855) have been used to describe why continents and mountains are higher than the surrounding ocean basins. Studies by Hayford and Bowie (1912), Woollard (1969), Dorman and Lewis (1970 a, b), Simpson and others (1986), and others considered the relationships among topography, gravity, and isostasy. Niskanen (1939), Gutenberg (1941), and others showed that unloading of large areas of North America and Europe, following the melting of continental ice sheets, resulted in isostatic rebound of the region.

The "backstripping" technique of Watts and Ryan (1976) in conjunction with lithospheric flexural modeling allowed them to model loading of the Gulf of Lion through time. In this gulf, the subsidence rate was exponential which suggested contractional cooling as an important non-sediment load "driving force" for subsidence. Nunn and others (1984) applied this technique to the Michigan basin to develop subsidence and thermal histories of the basin. Both of these studies used flexural modeling for areas that are much smaller than the areas of continental glaciation, but they are still larger than the Nashville dome.

Isostatic adjustment to Holocene deglaciation is well documented, and data obtained from these studies were used in modeling the viscosity and structure of the mantle (McConnell, 1968; Cathles, 1975; and others) and the lithosphere (Brotchie and Silvester, 1969; Walcott, 1970; Peltier, 1984; and others). Relatively large areas were involved in glaciation and a short period of time was involved during both glacial loading and inter-glacial unloading of the crust. Models of this isostatic response yield estimates of mantle viscosity of about 10^{21} Pa s and an effective elastic lithosphere of varying thicknesses, e.g. about 30 km (Brotchie and Silvester, 1969) to over 200 km, (Peltier, 1984), or a lithosphere with its thickness modified by zones of weakness within it (De Rito and others, 1986). The importance of the elastic properties of the lithosphere relative to the fluid properties in the asthenosphere increases with the effective thickness of the lithosphere; however, as the area undergoing isostatic adjustment increases, the importance of the elastic lithosphere decreases.

Although the Nashville dome is not large relative to the area of a continental glacier, it still may be large enough for isostatic adjustment. The isostatic response to the unloading of Lake Bonneville provides an example on a scale comparable to the Nashville dome (Figure 3). Lake Bonneville has an area of about 50,000 sq km; whereas, using the sea level contour, the area of the dome is about 40,000 sq km. The smaller Central Basin is about 14,000 sq km, but the basin has lost by erosion a mass of rock that is almost comparable to the mass of water removed over the larger area of Lake Bonneville.

The shoreline features associated with the Bonneville Stage (15 Ka) of Lake Bonneville are warped as the result of isostatic adjustment following the decrease in lake loading. Elevations of these strand lines allowed Crittenden (1963a, b) to calculate a relaxation time (time for 63% of the potential isostatic adjustment to be accomplished) of about 4,000 years and to estimate a viscosity of about 10^{20} Pa s at compensation depth for that region. The most recent and sophisticated modeling of Lake Bonneville by Bills and May (1987) confirmed Crittenden's viscosity estimate and it provided constraints on the thickness of the elastic

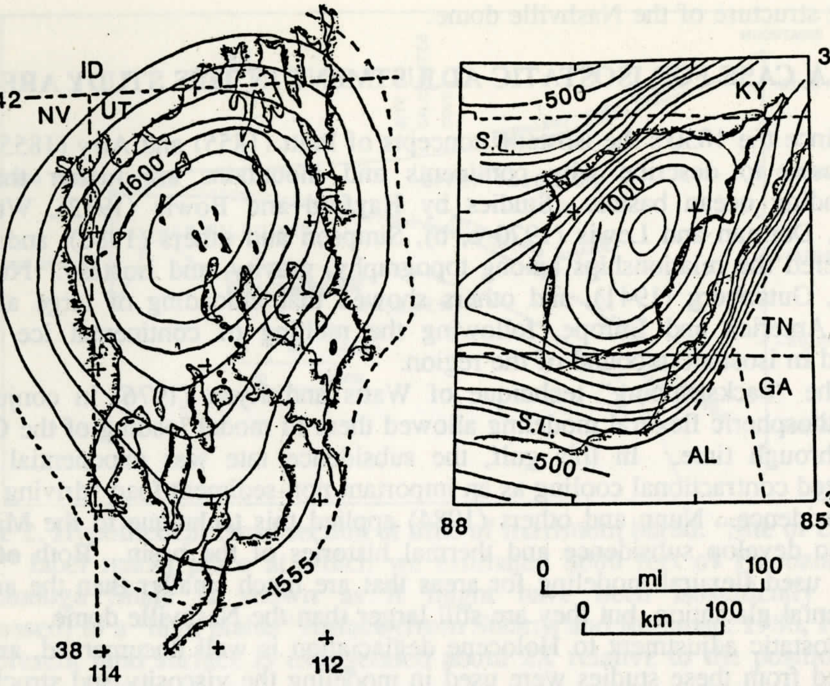


Figure 3. Approximately equivalent contour patterns for the isostatic uplift at Lake Bonneville and the proposed isostatic uplift of the Nashville dome. Both areas are at the same scale. The contour interval for Lake Bonneville is 10 meters with the contour pattern based on least squares best fit after removal of "structural" highs and lows, which allows the uplift pattern to be extended around the margin of the basin. The structure of the dome is based upon a 15-minute average on the base of the Chattanooga Shale and contour interval above sea level is 100 feet (30 m). The spacing of spot elevations at Lake Bonneville is between 10 and 15 minutes throughout the basin. The contour interval of 10 m for water verses "30 m" for the rock (density 2700 kg/m^3) are essentially loading equivalents.

lithosphere (23 ± 2 km) and the base of the asthenosphere (> 300 km). Lake Bonneville is in the Basin and Range Province, where the lithosphere has been thinned through extension. As a result, the crust is probably thinner and weaker in the Lake Bonneville area than in central Tennessee and the underlying asthenosphere at Lake Bonneville is about an order of magnitude more fluid than was calculated from the rates of post-glacial rebound of North America and Fennoscandia (10^{21} Pa s, Cathles, 1975). Applying the same equation that Crittenden (1963a, formula 7, p. 5526) used to calculate the relaxation time for Lake Bonneville, but substituting dimensions of the Central Basin and increasing the viscosity of the compensating layer to 10^{21}, a relaxation time of about 90,000 years was calculated for the Central Basin. This value is 22 times longer than calculated for Lake Bonneville, but time is not critical because unloading of the Central Basin is slow and is measured in meters per million (not thousand) years. A relaxation time approaching a million years could still provide an isostatic response.

Because the warped and raised remnants of the Lake Bonneville strand line provide an ideal setting for studying isostatic adjustment, the same simple approach to isostatic adjustment as used in this study was applied to Currey's (1982) 181 documented Bonneville Stage, shoreline elevations (Reesman, and others, in manuscript). Lake water loading was estimated at each of Currey's sites, by averaging water depths over a wide range of areas (22.5 to 120 minutes) using unweighted rectangular areas (as used in this study) and cosine-weighted. circular areas with diameters of 97.5 to 120 minutes. The relationship between load and elevation was evaluated through least squares regression for each of the areas of load averaging. This statistical technique allowed us to calibrate the isostatic effects of this very short term loading event in this specific setting (for additional information, see "Selection of a Best No-load Map"). Figure 3 compares the isostatically warped pattern of Bonneville Stage, shoreline features with a generalized structural contour map of the Nashville dome, which we believe to be the product of isostatic adjustment. The structure of the dome is based upon 15-minute averaging of elevations on the base of the Chattanooga Shale (see Figure 5B). The contours of the Lake Bonneville area were computed from least squares technique which allows us to extend the isostatic deflection beyond the edge of the lake basin. The elevation of the lake is calculated to have been 1551.6 ± 2.1 m. The contour interval of the lake area is 10 m (33 ft.); whereas, the contour interval of the Nashville dome is 100 feet (30 m). This three-fold difference in contour interval is roughly equal to the 2.7 difference in the ratio of densities of the rock and water. We have tried to make the two areas are as comparable as possible in scale, spacing of elevations, and density corrections.

Isostatic adjustment in response to erosion of the Central Basin seems reasonable by analogy with Lake Bonneville, especially if the great difference in the time for adjustment is considered. However, Walcott (1970) examined the effects of loading related to the Caribou Mountains in Alberta and found that this erosional remnant showed no effects of loading. These mountains are about the same size as the Central Basin but they are smaller than the Nashville dome. Thus, there is also evidence that adjustments on this scale will not occur.

ORIGINAL SMOOTHNESS OF THE PRE-CHATTANOOGA UNCONFORMITY

An originally smooth pre-Chattanooga surface will help in later evaluations of our calculated surfaces. Original smoothness of the pre-Chattanooga surface is suggested by the uniform shale thickness of 20 to 40 feet (6 to 12 m) and the lithology (widespread thin-bedded, black shale) of the shale over Northern Alabama, Tennessee, and parts of southern Kentucky. In Western Kentucky the Chattanooga or New Albany thickens to over 400 feet in the Illinois Basin, and east of 84 degrees in Kentucky the Chattanooga equivalent, the Ohio Shale, thickens to over 1700 feet going into the Appalachian basin. The Chattanooga Shale may have been deposited in relatively deep water (Rich, 1951, and Byers, 1977) or shallow water (Conant and Swanson, 1961, and Rheams and Neathery, 1984). Whether shallow or deep, the Chattanooga was deposited in a marine environment under quiet anaerobic conditions. The uniform thickness and lithology of the Chattanooga over such a large area helps to re-enforce our belief that whatever the depth, the surface was smooth.

CALCULATION OF ISOSTATICALLY ADJUSTED STRUCTURE

We assume that the pre-Chattanooga unconformity was in isostatic adjustment at the time it was covered by Chattanooga dark mud. We also assume that the present land surface (an active unconformity) is now in isostatic adjustment. Our goal is to calculate a surface for the pre-Chattanooga unconformity that has been isostatically adjusted for mass eroded from below the Chattanooga in the Central Basin, and the mass weighing it down elsewhere. The resulting surface can be described as an isostatically adjusted no-load structure of the Chattanooga. We will refer to these surfaces as "no-load surfaces".

To calculate elevations for no-load surfaces in areas like the Highland Rim, where Chattanooga and post-Chattanooga rocks lie above the unconformity, the load above the unconformity is mathematically "removed" and the surface adjusted upwards in a theoretical isostatic response to this unloading. The density used for the "removed" rock was 2.7 g/cm^3 (close to the value used for Bouguer gravity calculation of 2.67 g/cm^3 and very close to the density of limestones measured in quarries in central Tennessee) and the density used at compensation depth was 3.3 g/cm^3. Thus, for each foot of rock removed above the Chattanooga, the no-load surface rises 2.7/3.3 or 0.8181 feet above the existing structural surface. Within the Central Basin where the Chattanooga surface projects above the land surface, rock with a density of 2.7 g/cm^3 was "added" up to the projected base of the Chattanooga surface, so the no-load surface subsides by 2.7/3.3 or 0.8181 feet for each foot of added strata.

AREAS OF ISOSTATIC COMPENSATION

General

Isostatic adjustments take place over an area of compensation, and the minimum size for this compensated area must be determined. To proceed, elevations of the present land surface are needed that are averaged over a specific area. We used one or more 7.5 minute quadrangles, which will be referred to as "blocks", with the average elevation of the block attributed to its center. By using blocks with averaged elevations the crustal load (load is a function of elevation) is spread in an attempt to correct for the load spreading characteristics of the elastic lithosphere. Structural elevations of the Chattanooga were also averaged over blocks, but as will be shown later this may not be necessary for calculation of the no-load surface.

For convenience, average elevations of 7.5-minute quadrangles were used to start the process of isostatic adjustments. Because we are using a "trial and error" approach to obtain the appropriate block size, we wanted to start with a block size that is too small, and end with a size that is probably too large. A 7.5-minute block is roughly equivalent to spreading the load over a 7.0 km radius. Larger block sizes were formed by averaging groups of 7.5-minute blocks, which produces groups with a correspondingly larger effective radius. For example, a group of thirty-six, 7.5-minute blocks makes one 45-minute rectangular block (about 82. by 68. km) that has an approximate effective radius of 42 km. The radius is equivalent to the "radius of regionality" used by Vening Meinesz in his "regional isostatic system" (Heiskanen and Vening Meinesz, 1958, p. 137).

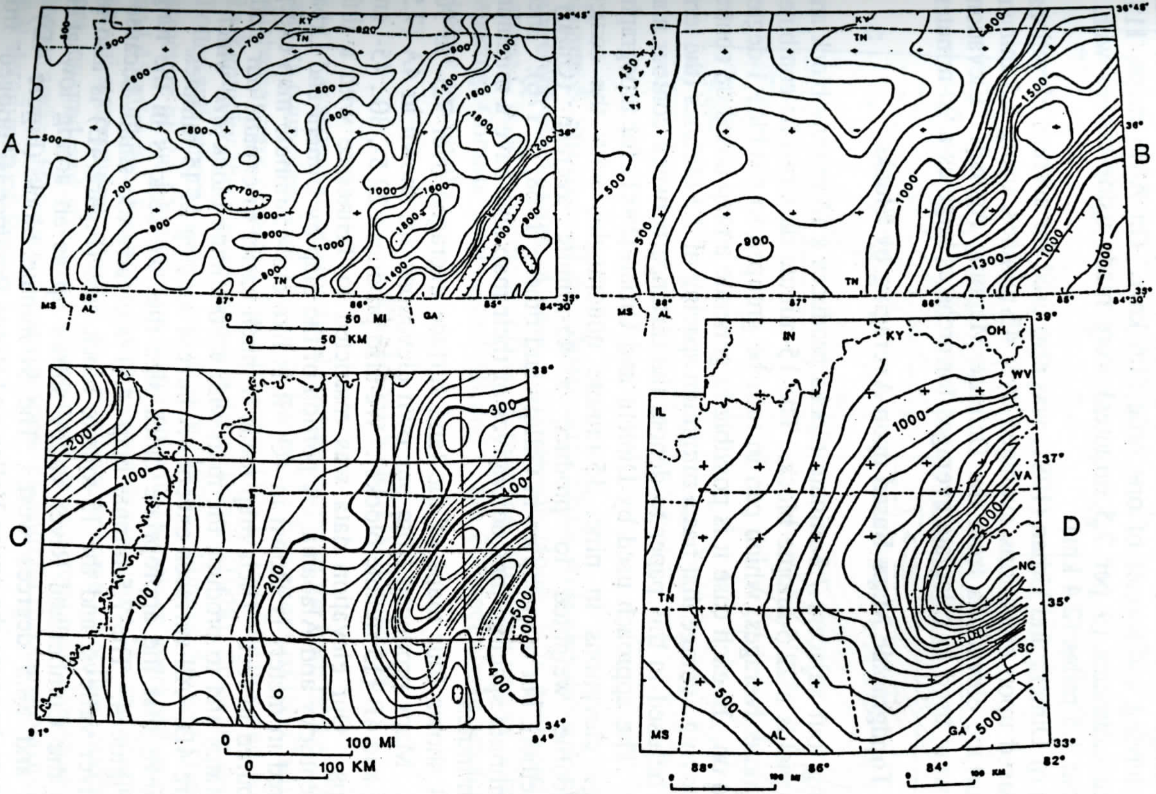


Figure 4. Contour maps of the averaged land surface elevations. Figures 4A and B are limited to Tennessee because they were made with our limited land surface elevations. 4A has a 7.5-minute block size and the block size of 4B is 15 minutes, both of these maps have 7.5-minute spacing. Figure 4C is a portion of the 45-minute land surface map of Diment and Urban (1981), which is contoured in meters. Their map made it easy to expand the averaged land surfaces into surrounding states. Figure 4D is a 90-minute block map made by combining the 45-minute blocks of Diment and Urban. Contour interval is 100 feet except in 4A and B, in which the interval above 1000 is 200 feet.

Estimating the Average Land Surface Elevations

Average land-surface elevations were determined from 7.5- and 15-minute topographic quadrangles and AMS 1 x 2 degree topographic quadrangles.For 7.5-minute topographic maps the average elevation was determined from point elevations at 81 regularly spaced points. For 15-minute quadrangles, 64 regularly spaced local average estimates were made (16 estimates per 7.5 minutes), each within a circle having a diameter of one mile (1.6 km). On AMS maps, 1152 regularly spaced estimates (9 per 7.5 minutes) were made using circles with a diameter of about 1.5 miles (2.4 km).

The map of Diment and Urban (1981) was also used for average elevations. Their map is based upon elevations averaged over 45-minutes, using 2.5 minute, average land-surface elevations from the Defense Mapping Agency. Elevations from this map were used to calculate averages for areas as large as a 90-minutes.

Topographic Maps Based upon Averaging by Blocks

Averaging four adjacent 7.5-minute blocks produces a 15-minute block and averaging nine yields a 22.5-minute block. The 15-minute data can be combined to create 30-minute averages, which can in turn be grouped to produce 1-degree averages and so on. In each case it is possible to generate a map based on control points with a moving average that has a minimum spacing of 7.5 minutes (the term "spacing" will be used in this paper to denote the distance between contour map control points). The approach used by Diment and Urban (1981) for topography was to average elevations in nine, 15-minute quadrangles with the center-quadrangle, double weighted to produce a 45-minute average (Gilluly's unpublished method, but map shown in Gilluly and others, 1968, p. 176). Their map has a 15-minute spacing, thus derivatives of their map also have a minimum spacing of 15 minutes.

Figure 4 shows four maps of the land surface averaged over 7.5- to 90-minute blocks. Maps produced from our own elevations are shown in 4A (7.5-minute blocks) and 4B (15-minute blocks). We also produced 22.5-, 30-, 45-, and 60-minute maps. Our elevation data was restricted to Tennessee with a little overlap into Kentucky and Alabama. A portion of the map by Diment and Urban (1981) contoured in meters is shown in 4C. It is like our 45-minute map except ours was contoured in feet. Their map made it much easier to expand our study into adjacent states and to produce the map with a 90-minute block size which is shown as Figure 4D. All surfaces except 4A have a spacing of 15 minutes.

We believe that the appropriate block size that we seek is in the range illustrated in Figure 4. The 7.5-minute map (4A) is clearly too small, because it shows small river valleys, and the 15-minute block size (Figure 4B) is also too small because the Cumberland River valley shows up as an arcing low at 36 degrees north and 88.5 degrees west. The 90-minute block (Figure 4D) is probably too large because the entire Valley and Ridge Province is "bridged" from the Cumberland Plateau to the Appalachian Mountains. A block in the range larger than 15 minutes and smaller than 90 minutes should be about right, maybe 45-minutes like Figure 4C.

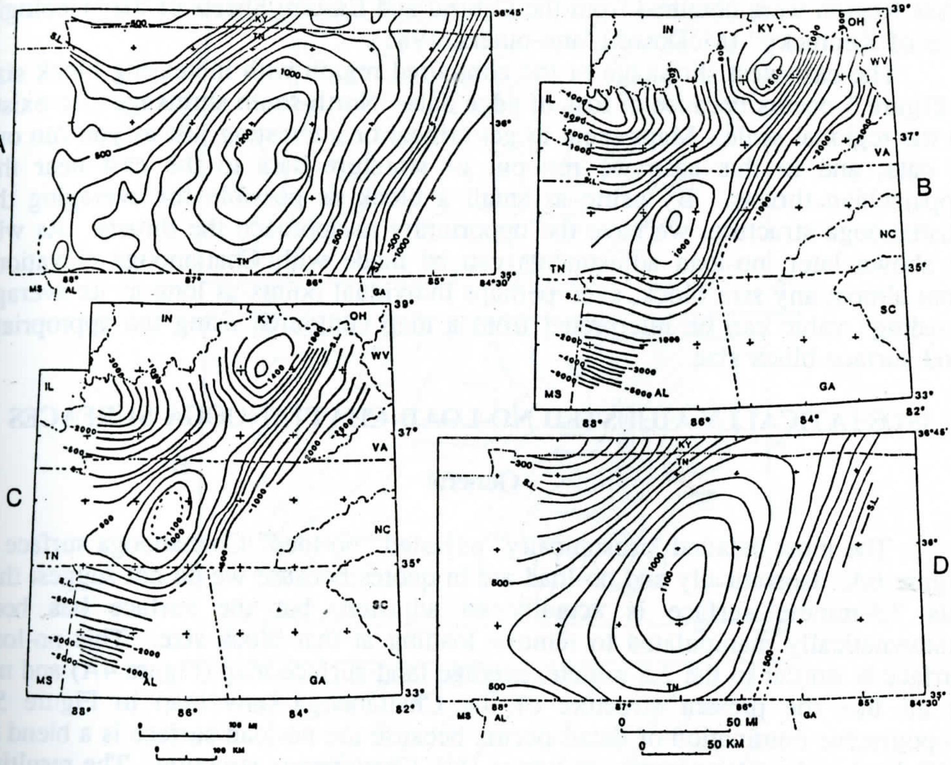


Figure 5. Structure of the Chattanooga Shale with elevations averaged using four different block sizes. A uses a 7.5-minute block; B is averaged over 15-minute areas; C is 30 minutes; and D is 45 minutes. All maps except A have a 15-minute spacing. Contour intervals of 5B and C are 200 feet above sea level and 500 feet below sea level.

PRESENT STRUCTURE MAPS

A sequence of structural contour maps on the base of the Chattanooga Shale and its equivalents are shown in Figure 5. Elevations of the Chattanooga Shale in Alabama were obtained from Rheams (1981) and Rheams and Neatherly (1984). In Kentucky the elevations were from Potter (1978), Schwalb and Potter (1978), Potter and others (1984), and Fulton (1979). Figures 5A and D show the present structure of the Chattanooga Shale only in Tennessee averaged using blocks of 7.5- and 45-minutes. Figures 5B and C were drawn using 15- and 30-minute blocks for the tri-state region. Notice that contours become smoother, maximum elevations are lowered and structural lows diminish as a result of averaging elevations over larger areas. However, all maps show essentially the same Nashville dome as the full detail map of Figure 1.

Because outliers of Chattanooga equivalent shale are not extensive in the Blue Grass Region of Central Kentucky, which is the erosional counterpart to the Jessamine dome, the structure on the base of the shale, shown in Figures 5B and C, was approximated by estimating the thickness of the various formations that had been removed by erosion. These approximations were made in 15-minute blocks. Stratigraphic thicknesses, outcrop patterns, and land-surface elevations in the Blue

Grass Region were obtained from the Central and Eastern Sheets of the "Geologic Map of Kentucky" (McDowell, and others, 1981).

The east-west shrinkage of the contoured region with increasing block size in Figure 5 occurs because of loss of edge data. North-south shrinkage also exists on the regional scale. Maps seem to get critical or interesting just as you run out of data, and in this case we run out of structure data to the east near the Appalachian thrusts. By using as small a block as possible for averaging the Chattanooga structure, we have the opportunity to approach the thrusts. As will be shown later, no-load adjustments can be made with Chattanooga elevations from almost any size block, even perhaps individual points as long as its average "loading" value can be interpreted from a map contoured using the appropriate land surface block size.

ISOSTATICALLY ADJUSTED NO-LOAD CHATTANOOGA SURFACES

General

The most detailed "isostatically" adjusted "no-load" Chattanooga surface is Figure 6A. Isostatically and no-load are in quotes because we do not suggest that this 7.5-minute surface is actually so adjusted, but the surface has been mathematically manipulated to remove loading at that block size. This no-load surface is similar to the 7.5-minute, average land-surface map (Figure 4A) and not at all like the present structure of the Chattanooga (any map in Figure 5). Topographic domination of detail occurs because the no-load surface is a blend of 82% land-surface topography and only 18% Chattanooga structure. The resulting no-load Chattanooga surface has too small an area of topographic averaging, as evidenced by the topographic details that it displays. It is shown only as a basis for comparison with smoother maps averaged over larger areas.

Figure 6B is based on a 22.5-minute block size for both the land surface and the Chattanooga structure. Judging from the slight expression of river valleys, this block size is still too small. Neither 6C, or D show topographic details, so they may have sufficiently large topographic block sizes. Figure 6C has a 45-minute, land-surface, block size and a 15-minute Chattanooga structural block. Land-surface, block size was increased to 60- minutes in Figure 6D to help bridge to the 90-minute surfaces shown in Figure 7.

Figure 7 shows two things: first, the general map pattern is unaffected by the size of the area over which the Chattanooga structure is averaged; second, smaller details can be seen within the similar looking pattern if structure is averaged over small areas. The maximum detail is shown on 6A in which midpoints of the 7.5-minute elevation blocks of Chattanooga structure are used. The minimum detail is seen on 6D in which 60-minute Chattanooga elevations are used. The east side of the pattern disappears on 7D because the map control points must be 30 minutes "inside" the edges of the structural data. If the pre-Chattanooga unconformity was a smooth uniform plain, we might assume that the surface may have been in "better" isostatic adjustment than our present very active isostatic surface. When the pre-Chattanooga surface was buried by later sedimentation, it was sandwiched into the stratigraphic sequence to record the effects of both isostatic subsidence and non-isostatic structural events. The smoother the Chattanooga was originally, the better the non-isostatic events might be discerned, if isostatic movements can be removed as we are attempting to do.

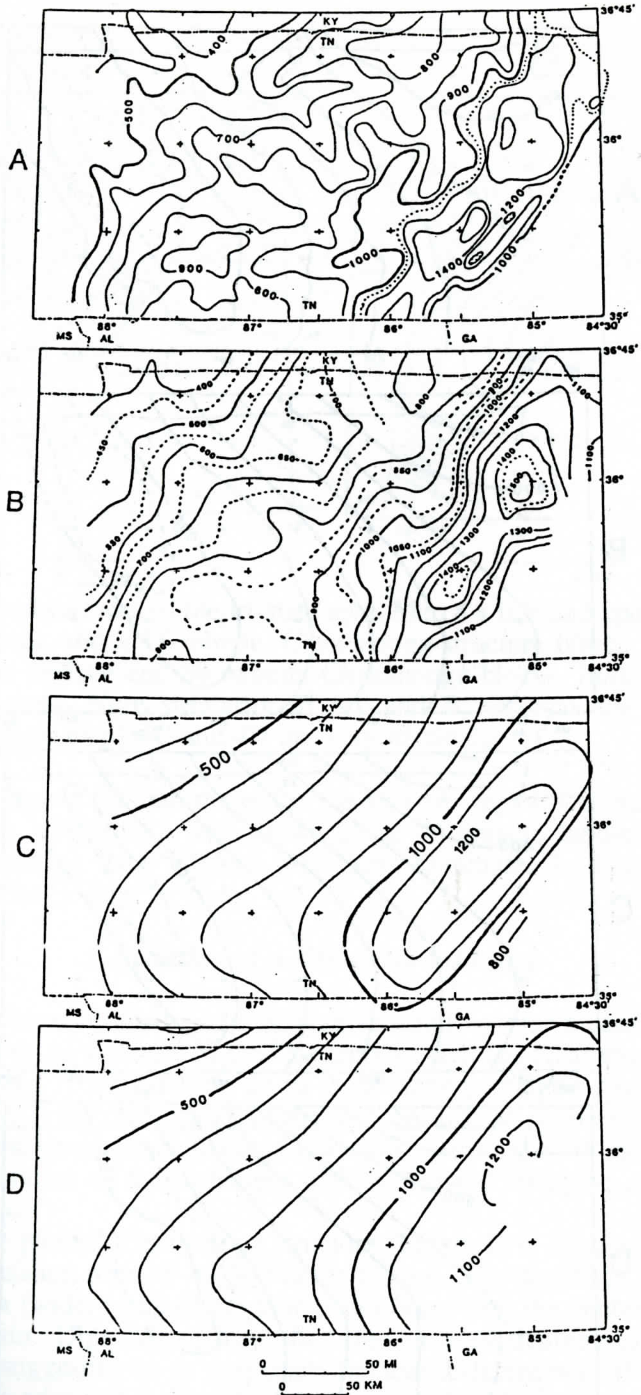


Figure 6. No-load maps on the base of the Chattanooga Shale. A has a 7.5- minute block size for both land surface and structure; B has a 22.5-minute block size for both the land surface and structure; C has a 45-minute land surface and a 15-minute structure block; and D has a 60-minute land surface and a 30-minute Chattanooga block size. Spacing of A and B is 7.5 minutes, all others have 15-minute spacing. Contour interval of 6A is 200 feet above the 1000-foot elevation.

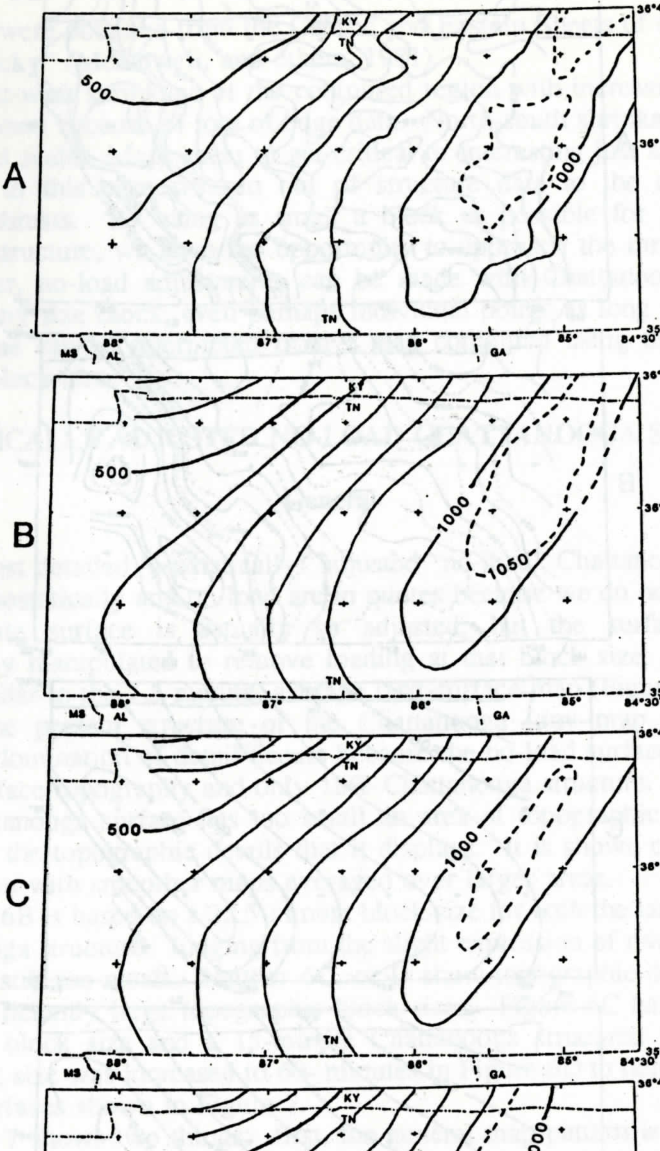


Figure 7. Four versions of no-load Chattanooga structure using 90-minute land surface blocks. A has 7.5-minute structure block size; B has 15 minutes; C is 30 minutes; and D is 60 minutes.

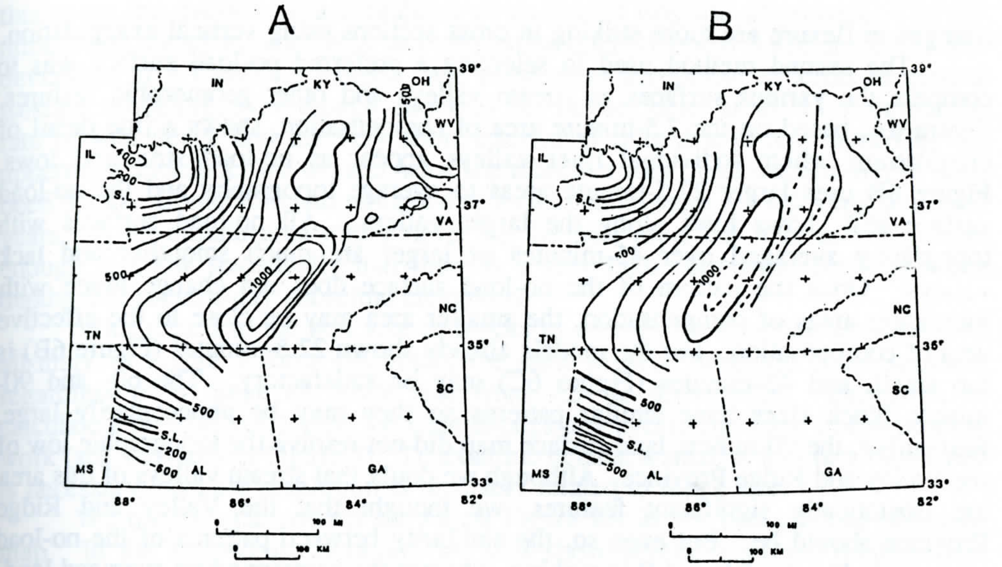


Figure 8. No-load surfaces of the tri-state area. Map 8A is based upon a 45-minute, land-surface block and a 15-minute, Chattanooga-structure block. Map 8B has a 90-minute, land surface and 30-minute, Chattanooga blocks. Both A and B have 15-minute spacings. Note similarity of 8A and B, whereas the corresponding land-surface, block maps 4C and 4D are very dissimilar.

Figure 8 shows no-load maps for the tri-state region (Alabama, Kentucky, and Tennessee). Figure 8A was drawn using 45-minute, land-surface and 15-minute, Chattanooga data. Figure 8B uses 90-minute, land-surface and 30-minute, Chattanooga data.

Selection of a "Best" No-load Map

Our preferred no-load surfaces were selected by four means; 1) looking at the present full detail structure of the Chattanooga near the escarpment between the Central Basin and Highland Rim, 2) by a cut-and-try method relative to surface geomorphology, 3) the relationship of the no-load surfaces with existing structural, stratigraphic, and geophysical features (this will be discussed in more detail later) and 4) by calibration of the technique with the isostatic warping of shorelines in Lake Bonneville.

First, the present Chattanooga structure (Figure 1) is itself suggestive of an area of compensation over 25- to 30-minute blocks. The dome has a broad nearly flat crestal area (widely spaced contour lines) occupying the center of the nearly flat Central Basin. Dips steepen away from the crest, particularly to the northwest and southeast, suggesting to us that sharp flexural differences of the Chattanooga surface near the edges of the basin might have resulted from isostatic response to effects of loading under the rims and the removal of pre-Chattanooga within the basin that occur at the 300 to 500 foot (98 to 164 m) escarpments between the basin and rim. An abrupt flexure shift is distributed 10-15 minutes inside the escarpment (at the edge of the basin), and maybe another 10-15 minutes outside the escarpment (beneath the edge of the rim), suggesting a region 25-30 minutes wide over which much of the supposed isostatic compensation may be occurring. These changes in flexure are more striking in cross sections using vertical exaggeration.

The second method used in selecting a preferred no-load surface was to compare the various surfaces to stream valleys and other geomorphic features. Figure 6A, based on the 7.5-minute area of compensation, shows a fine detail of crenulations where individual river valleys appear as no-load structural lows. Figure 6B uses larger 22.5-minute areas to average topography and the no-load surface still shows lows along the larger valleys. All no-load surfaces with topography averaged over 45-minutes or larger are much smoother and lack valleys. Once the pattern of the no-load surface does not change much with increasing areas of compensation, the smaller area may be close to the effective area of compensation. For the models already shown 22.5-minutes (Figure 6B) is too small, and 45-minutes (Figure 6C) may be satisfactory. The 60- and 90-minute block sizes have similar patterns so they may be unnecessarily large. Remember, the 90-minute land surface map did not resolve the topographic low of the Valley and Ridge Province. Although we doubt that stream valleys of this area are isostatically significant features, we thought that the Valley and Ridge Province should be. But even so, the similarity between patterns of the no-load surfaces in Figures 8A and B is striking, whereas the corresponding averaged land-surface topography, Figures 4C and 4D are very dissimilar.

The technique of using elevations averaged over various areas as an approximation of loading was applied to old Lake Bonneville (Reesman, and others, in manuscript). Water depths were estimated throughout the lake basin at 2.5-minute intervals and the depths were averaged over various areas to provide a series of loads. Eleven rectangular areas were studied in detail with block sizes ranging from 45 to 120 minutes along with five cosine-weighted, circular areas. The average values were computed (not done by hand) with 7.5-minute spacings over the whole lake basin. The grided data were used to interpolate the water depth at each of Currey's (1982) 181 sites on the Bonneville Stage shoreline. The elevation of each site (X) was paired with the average water depth at that site (Y) and the X, Y, data were processed by least squares linear regression.

Residuals from the least squares processing were evaluated relative to the geology, geophysics, and geographic distribution in order to locate areas where post-Bonneville (15 Ka) neotectonism may have overprinted the isostatic adjustment. After refinement, the 82.5-minute area produced the best fit for the unweighted rectangular areas with areas from 67.5 to 97.5 yielding very good fits. The regression line has a slope of 3.41, an intercept of 1551.6 m, a correlation coefficient of 0.994, and a deviation of ± 2.1 m (Reesman, and others, in manuscript). This deviation is roughly equivalent to the estimated error of about ± 2. m assumed for the initial input X-Y data.

Currey and Burr (1988) suggest that half of the 240 feet (73 m) of isostatic subsidence in the deepest part of the lake basin occurred at a rate of about 7 cm/year within the interval of 15,500 to 15,000 ya and that the highest shoreline developed in a very short period of time, perhaps a few 10s of years. There model suggests that the basin had not subsided fully prior to a sequence of rapid fluctuations and catastrophic failure of the lake's threshold.

If subsidence was about 10 feet (3 m) from reaching its maximum (96% complete), the 75-minute or possibly even the 67.5-minute areas would have produced the best fit. Correlation coefficients of 0.994 and 0.991 and deviations of ± 2.22 and 2.72 meters were obtained from these areas without any corrections. These values are very close to the values obtained from the 82.5-minute area and this opens the question of acceptability of an area of isostatic adjustment. The 60-minute and the 45-minute areas had correlation coefficients of 0.988 and 0.978 and deviations of 3.1 and 4.3 meters. Are these areas acceptable in terms of general isostatic modeling, even though they may not be the "best" area? Considering the rate of subsidence and the very short interval of time that is represented by the uppermost shoreline, we wonder what the effective area of adjustment might have been if Lake Bonneville had remained at its maximum stage for a several thousand years. A 45-minute area might never be the best area, but it is certainly large enough to show statistically that there is a very high correlation between water depths and present-day shoreline elevations.

With regard to the best size of blocks to average the Chattanooga structure, all maps in Figure 5 show essentially the same present Nashville dome structure regardless of a range of block size from 7.5- to 45-minutes. Therefore, to calculate a no-load surface on the Chattanooga any averaging block size appears to work (see, surfaces in Figure 7). To examine details, a small a structure block size is necessary, because resolution is inversely related to block size.

We propose the pattern shown on Figure 8A as the best no-load structure map. In the following discussion we will show several geologic relationships that could be related to the no-load surfaces, all of which appear to fit the 45-minute (Figure 8A), no-load surface better than the 90-minute surface (Figure 8B) even though these two maps are similar.

Discussion of No-Load Surfaces

General: The regional no-load surfaces in Figures 8A and/or B will serve as the base for this discussion. These surfaces will be examined relative to the existing structure on the Chattanooga, general structural and stratigraphic considerations, and geophysical patterns in an attempt to build a case, based upon circumstantial evidence, for isostatic adjustment at the relatively small scale proposed in this study.

The Nashville Dome: None of the no-load Chattanooga surfaces presented herein, or any other surfaces or cross-sections with areas less than 150 minutes show any hint of the Nashville dome. Cross-sections (not presented herein) included different, but reasonable, densities for the surface rock and the compensating medium. If all reasonable isostatically adjusted no- load surfaces relating topography to structure remove the structure of the Nashville dome, then we conclude that the Nashville dome is the result of provincial isostatic adjustment to topography.

The general pattern of regional denudation, which is dominated by the retreat of the Cumberland Plateau escarpment, led to a modest broad pre- emergent ancestral Nashville dome. We infer that isostatic response to the current three-fold difference in the rates of denudation between the rim- forming Fort Payne Formation and the highly soluble Ordovician limestones within the Central Basin is responsible for the present location and shape of the crest of the dome. We believe both the basin and the dome are growing together. Because it has been only about 5 million years since the breaching of the Fort Payne Formation to form the Central Basin (Reesman and Godfrey, 1981), the major growth of the crestal portion of the dome is also about 5 million years old.

Isostatic growth of the dome implies that for every 150 feet (46m) of erosion (stratigraphic denudation) that occurs in the center of the basin each million years the land surface is lowered by 27 feet (8 m). Rocks as yet uneroded rebound 123 feet (38 m) and the structural amplitude of the dome is increased by that amount.

The Jessamine Dome: Unlike the Nashville dome, the Jessamine or Lexington dome of Central Kentucky was resolved on the no-load surfaces in Figures 8A and B. Outliers of New Albany (Chattanooga) Shale are not preserved within the Blue Grass Region as they are in the Central Basin so the structural surface on the shale was estimated over the Jessamine dome. If the uncertainty associated with our estimates of the elevations over the crest of the dome is 100 feet (30 m), the uncertainty on the no-load surface would be 18 feet (5 m). It would take an error of between 700 to 800 feet on the approximated base of the shale to remove the 150 feet or so of closure on the no-load map. Thus, we believe Figures 8A or B provide an acceptable configuration of the no-load surface over the dome. Even though the Jessamine dome appears on the no-load surfaces, it must have a large isostatic component because it is about the same size as the Nashville dome.

Differences Between The Nashville And Jessamine Domes: Differences between the Nashville and Jessamine domes are not limited to their expression on the no-load surfaces. Erosional outliers and scarps of Mississippian carbonates were buried by Pennsylvanian sediments and preserve a record of Late Mississippian or Early Pennsylvanian uplift on the Jessamine dome (Wanless, 1975). There are no known erosional scarps that have been buried by younger sediments in or around the Nashville dome to indicate a definite period of partial uplift as there is around the Jessamine dome. The Silurian outcrop pattern on the pre-Chattanooga erosion surface was once thought to be related to the emergence of the Nashville dome. However, this Silurian belt runs north-south along the western side of the dome, so the western side of this high is constrained, but the eastern side is not. We believe the pre-Chattanooga high was not centered on the present Nashville dome.

The degree of faulting is very different on these domes. Normal faults with large (100s of feet) displacements are common on the Jessamine dome, whereas even minor faults with a few feet of displacement are rare on the Nashville dome. It seems logical to us that a structurally controlled Jessamine dome might have more faulting than the Nashville dome if it was produced by isostatic adjustment.

Cumberland Plateau Dome: All no-load surfaces and cross sections consistently display an elongated high to the east of the Nashville dome under the Cumberland Plateau, an area that is now structurally low. The height of this dome is sensitive to block size, e.g. the no-load surfaces in Figures 6 and 7. Just as certainly as the Nashville dome's emergence resulted from isostatic adjustment, the absence of the "Cumberland Plateau dome" on the present-day, Chattanooga structure maps (Figure 5) is the result of isostatic suppression.

Because the Cumberland Plateau dome is partly under the Cumberland Plateau and the no-load surface is a blend of 82% topography and 18% structure, one might suspect that this dome is an artifact of the topography. However, the dome does not follow topographic trends as well as would be expected if the contours were only an artifact of topography.

The southern end of the Cumberland Plateau dome is located just west of

Appalachian thrusts and Sequatchie anticline, and it parallels them. This geographic relationship might suggest that the dome is a foreland bulge resulting from the thrustal loading on the eastern side of the Cumberland Plateau (as Quinlan and Beaumont proposed as an origin of the Nashville dome in 1984). However, as will be shown below, we believe this dome and the western margin of thrusting are basement controlled. The dome probably buttressed thrusting in Tennessee, but it was not formed in response to thrusts. However, thrusting could have enhanced the structure.

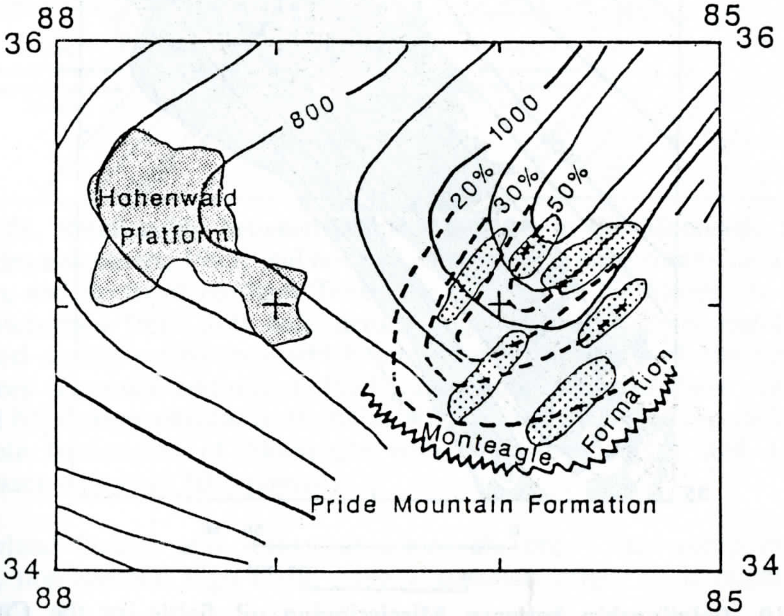


Figure 9. Relationship between no-load surface and oolitic shoals in the Upper Mississippian Monteagle Limestone and the Hohenwald platform. The orientation and distribution of oolitic tidal bars (from Handford, 1978, Figure 11) and the contours of oolite isoliths (from Handford, 1978, Figure 6) produce a very interesting pattern relative to the no-load contours. Handford's interpretation of the facies relationship between the Monteagle and the more clastic Pride Mountain Formation also fits the no-load contours. The Hohenwald platform was considered to have been an island during the deposition of the Chattanooga (Conant and Swanson, 1961).

It is unlikely that the Cumberland Plateau dome existed as a topographic high during deposition of the Chattanooga or was very large during deposition of the overlying Mississippian carbonate sequence, because there is little or no evidence for thinning of these sediments across the dome. The Mississippian carbonates do thicken toward the southern end of the dome where the abundance and distribution of oolite shoals in the Mississippian Monteagle Limestone (Handford, 1978) appear to be related to the dome as does the facies relationship between the Monteagle and its more clastic equivalent, the Pride Mountain Formation. Figure 9 shows the Handford's Monteagle patterns along with the no-load contours from Figure 8A.

On the northern end of the dome, the distribution of oil fields in the Mississippian Fort Payne and Monteagle Formations may be related to the dome.

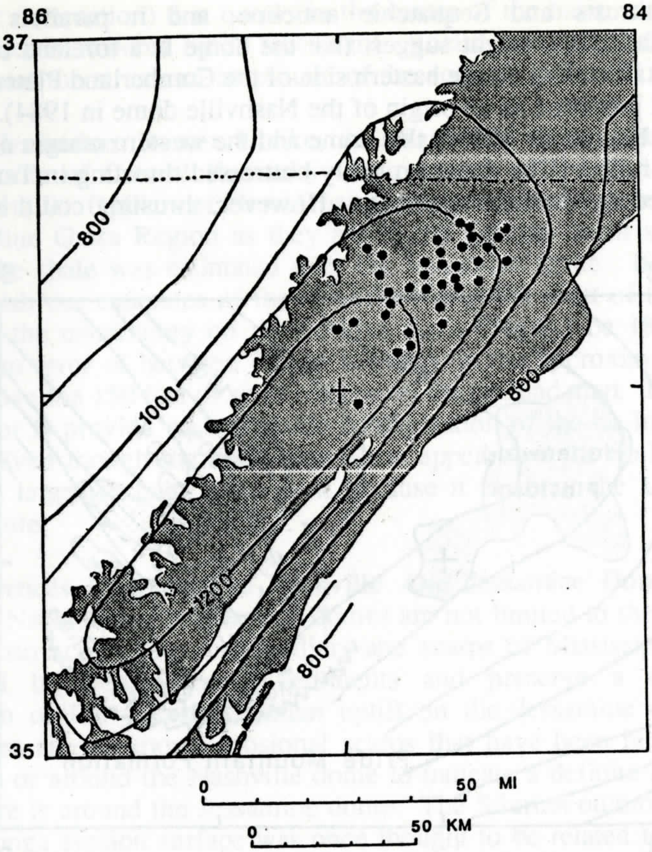


Figure 10. Relationship between Mississippian oil fields on the Cumberland Plateau and the no-load contours from Figure 8A. These fields are producing from either the Fort Payne (mud mounds) or Monteagle (oolitic zones) Formations. The stippled area corresponds to the Pennsylvanian capped Cumberland Plateau.

Figure 10 shows the relationship between the no-load surface of Figure 8A and the general position of these oil fields. The stippled area is the Pennsylvanian capped Cumberland Plateau. Although no-load contours generally conform to the topographic high of the Cumberland Plateau in Tennessee, northward into Kentucky they do not.

Figure 11 is a close-up of the northern end of the Cumberland Plateau dome to show how the configuration of the dome from Figure 8A appears to match the porosity trends in the oolitic shoals studied by Lumsden and others (1983) in the Monteagle Limestone. They concluded that these oolitic shoals tended NNE perpendicular to the shoreline which would have been to the east. Figure 11A shows the aggregate thickness of Monteagle with over 4% porosity (Figure 6 of Lumsden and others, 1983) and 11B shows the aggregate thickness of Monteagle with over 10% porosity (Figure 7 of Lumsden and others, 1983).

Relationship Between The Jessamine And Cumberland Plateau Domes: The 1000-foot contour lines in Figures 8A and B suggest a saddle between the Cumberland Plateau and Jessamine domes. But, the saddle is slight; its crest occurs at about 970 feet (295 m) on both 8A and B, so for practical purposes, the

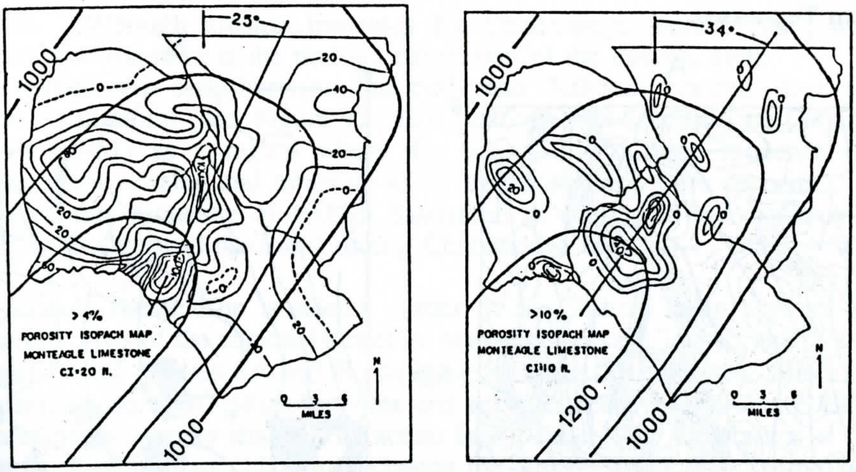


Figure 11. Relationship between porosity patterns in the Monteagle Limestone (Lumsden and others, 1983) and no-load contours in a three county area (Fentress, Morgan, and Scott) of northern Tennessee. Porosity in the Monteagle Limestone was determined from drill hole density logs. Porosity occurs mainly in spar cemented oolitic grainstones which are repeated as parts of shoaling upward sequences (Lumsden and others, 1983). They concluded that these marine shoals trended NNE perpendicular to shoreline which was to the east. Figure 11A shows aggregate thicknesses of Monteagle with over 4% porosity and 11B shows thicknesses with over 10% porosity.

Cumberland Plateau and Jessamine domes are one. This composite dog-leg shaped, non-isostatic high could have a common origin or it might be more complex.

The coincidence between the East Continent gravity high (Keller and others, 1982) provides evidence for a genetic relationship between the Cumberland Plateau and Jessamine domes and suggests these no-load highs could be related to basement structure. Figure 12A shows a portion of the Bouguer gravity anomaly map of Keller and others (1980) with the 45-minute, no-load contours superimposed. The 45-minute map provides a better fit than the 90-minute surfaces in terms of location and orientation between the two. This gravity and magnetic high was modeled by Keller and others (1982) as a rift related feature. The feature also appears as an isostatic residual gravity high (Simpson and others, 1986). The magnetic patterns (Zietz and Bond, 1984) also relate to the Cumberland and Jessamine domes. They are not so striking and are too complex to show contour lines at this scale. Figure 12B shows areas with total magnetic intensities of over 1800 gammas, and the location of the regional maximum intensity of 3600 gammas marked by an "X" near the center of the Cumberland Plateau dome. In general, areas of high magnetic intensities decrease northward toward the Jessamine dome but extend through the dog-leg connection. Figure 12B also shows the major normal surface faults associated with the Jessamine dome and the Rome trough and the major thrusts of the Valley and Ridge.

Paleocurrent directions in the southern portion of the Corbin Member of the Lee Formation (L. Pennsylvanian) parallel the trend of the no-load contours along the eastern flank of the Jessamine dome (Rice, 1984) and they are consistent with the trend reported by Jackson (1984) in the equivalent Rockcastle Conglomerate in northern Tennessee.

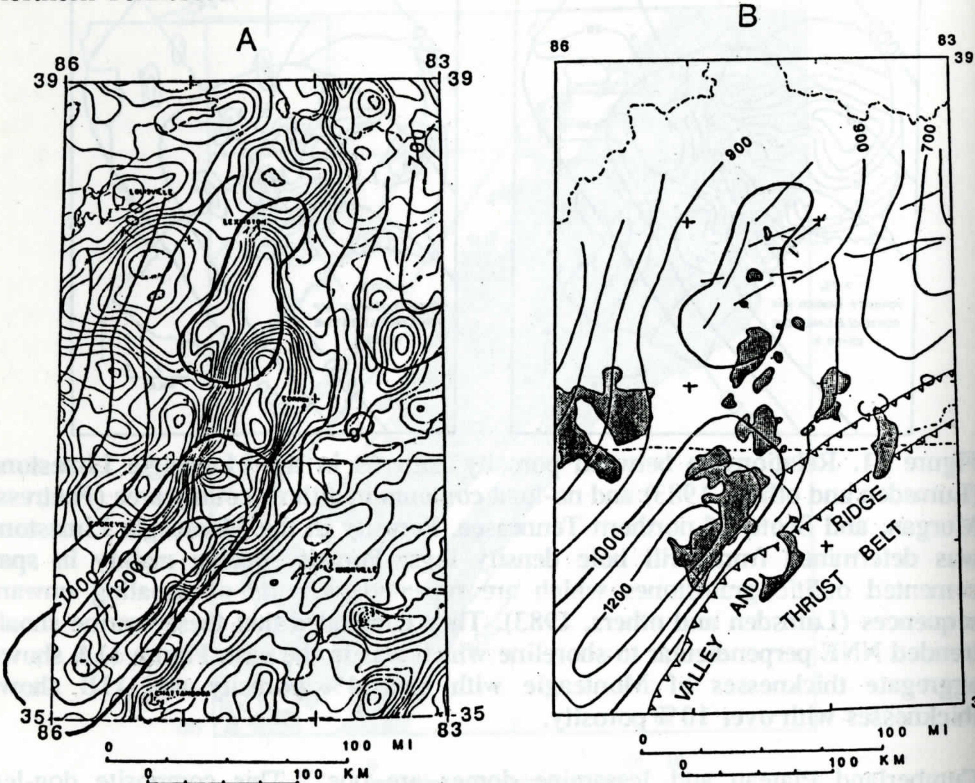


Figure 12. Evidence for basement control of the Cumberland Plateau and Jessamine domes. Figure 12A is a portion of the Bouguer gravity anomaly map of Keller and others (1980) with the no-load contours from Figure 8A superimposed. The gravity high between the 1000-foot contours of the two domes is the East Continent gravity high (E.C.G.H.). The E.C.G.H. was modeled by Keller and others (1982) as a rift related feature. The associated magnetic patterns are too complex at this scale, but the stippled areas in Figure 12B have magnetic intensities above 1800 gammas and the "X" marks the location of the maximum (3600 gamma) intensity (from Zietz and Bond, 1984). In addition to the no-load contours, the normal faults associated with the Jessamine dome and the Rome trough and the major thrusts of the Valley and Ridge are shown.

Clifton Saddle-Pascola Arch: The westward trending nose in Figures 6 and 7 appears to be a minor feature on these no-load surfaces, but when examined at the regional scale in Figures 8A and B, its significance becomes more apparent. This nose is the eastern extension of an arch that separates the Illinois Basin to the north from the Warrior basin to the south. A north-south cross section at about 88° yields a 250 mile (400 km) wide (basin to basin) arch with 1000-feet (300 m) of non-isostatic structure.

The western most arc of the 600-foot contour line occurs near Clifton, Tennessee and represents the non-isostatic component of the Clifton saddle, which Wilson (1939) interpreted as a structural arch between the Nashville and Ozark domes. Grohskopf (1955) found the main structural high buried beneath 2000 feet of sediments in the northern portion of the Mississippi Embayment near Pascola, Missouri. Although erosion truncates the Chattanooga west of 88°, we feel confident that the nose is the no-load expression of the Pascola arch.

Astride the arch between the 600- and 700-foot contour lines is the Hohenwald platform (see, Figure 9). This platform was described by Conant and Swanson (1961) as a region where the normal black shale sequence of the Chattanooga is missing and sands up to 20 feet (6 m) thick were deposited. They interpreted this sequence of Hardin Sandstone as sediments associated with an island (the Hohenwald platform) during Chattanooga time.

Illinois Basin: The southeast corner of the Illinois basin appears in the northwest corner of the regional structure maps (Figures 5B and C) and in the no-load surfaces of Figures 8A and B. Judging by shale thickness, the Illinois basin was subsiding during Chattanooga time and accumulated over 400 feet (122 m) of Devonian-Mississippian shale. The actual base of the New Albany lies at depths in excess of -4000 feet (1200 m) between the Rough Creek and Pennyrile fault systems. In contrast, the isostatically adjusted base is -400 feet (-120 m) using 45-minute averaging or -300 feet (-90 m) using 90-minute averaging.

Northern Portion Of The Warrior Basin: The base of the Chattanooga Shale lies at depths in excess of -6300 feet (-1920 m) in the Warrior basin. The isostatically adjusted equivalent depth is -500 to -600 feet (-150 to - 183 m) for the 90- and 45-minute, no-load surfaces (Figure 8A and B). The thin and relatively uniform character of the Chattanooga in the Warrior basin is much like that of southern Tennessee and suggests that at the time of deposition the conditions throughout northern Alabama and southern Tennessee were very similar.

Western Portion Of The Appalachian Basin: East of the Jessamine dome, the Chattanooga equivalent Ohio Shale thickens from about 100 feet (30 m) to 1000 feet (305 m) near 82° 30' longitude. In eastern most Kentucky, it reaches a thickness of about 1700 feet (520 m) and the base of the shale is about -4000 feet (-1220 m) (Fulton, 1979). To the northeast along the Virginia-West Virginia line, the thickness of late Devonian sediments is about 10,000 feet (3050 m) (Conant and Swanson, 1961). Subsidence and sedimentation were active in the east during Chattanooga time.

The isostatically adjusted no-load surface leading into the Appalachian basin in eastern Kentucky (Figure 8A and B) is unlike that leading into the Illinois and Warrior basins, in that there is little evidence of a non- isostatic low in eastern Kentucky. If the top of the shale had been picked for adjustment, the surface to the east of the Jessamine dome would appear even flatter because the top of the dome would be about 20 feet (6 m) higher, resulting from a 100-foot thickness, whereas the very eastern most contours in Figures 8A or B would be over 200 feet (61 m) higher and the difference in elevation between the dome and the eastern edge would be reduced from about 500 feet (152 m) to less than 300 feet (91 m).

Isostatic adjustment of the elevations on the base of the shale in and near the Middlesboro syncline from Potter, and others (1984) show that there is no strong bulge paralleling the Pine Mountain thrust block (see, Figure 12B). The Pine Mountain thrust is part of the same group of thrusts (including the Sequatchie thrust) that trends parallel to the Cumberland Plateau dome. Because the Cumberland Plateau dome does not continue ENE, paralleling the Pine Mountain block, but rather trends northward, it seems doubtful that the Cumberland Plateau dome is a foreland bulge. Rather, its coincidence with geophysical trends having a deeper source suggests control by basement structure.

DISCUSSION

All geologists believe in isostatic adjustment at some scale, and isostatic adjustments can account for the major fraction of vertical motion. The percentage of believers increases with area of adjustment, and we are admittedly operating at the lower limit of belief. It seems worthwhile to try to filter the isostatic component from the non-isostatic component, in an attempt to understand the vertical motions on stable platforms. No-load surfaces appear to be a reasonable method for such filtering.

No-load surfaces are relevant only insofar as they can or have the potential to explain the previously unexplained, or insofar as they lead to new explanations. The no-load surfaces suggest the Nashville dome was and is being formed by the processes of isostatic adjustment and the Cumberland Plateau dome is an isostatically suppressed structure, once connected to the present Jessamine dome. The no-load structure shows coincident relationships to structural features (Pascola arch and Hohenwald Platforms), trends of lithology (Mississippian oolites), and geophysical patterns.

Evidence to support the patterns of the no-load surfaces on the base of the Chattanooga is purely circumstantial. However, we believe that their coincidences strengthen the case that such adjustments can occur over an extended period of time to produce adjustment within a relatively small area, such as the Nashville dome and the Central Basin.

Maps of no-load surfaces can be refined by adding more topographic detail (decreasing the spacing from 15-minutes) and by using weighting factors (e.g., the cosine-weighted circular areas that were used for the Lake Bonneville study). For these improvements it will be necessary to abandon hand calculations in favor of the computer and to use the average land-surface elevations that are available from the Defense Mapping Agency. However, we believe the contour patterns of the relatively simple (and geophysically unsophisticated) no-load surfaces are significant because the same general contour patterns persisted over a wide range of block size adjustments, this similarity of pattern was also noted in modeling the Lake Bonneville water loads.

MAIN CONCLUSIONS

Over a period of a few million years, isostatic adjustment can perhaps occur over areas as small as 45-minute quadrangles and can more certainly occur over 90-minute quadrangles. The Nashville dome is a product of isostatic adjustment that was initiated by the general pattern of regional denudation. The more rapid erosion of the limestones within the Central Basin during the past five million years resulted in more localized isostatic adjustments that define the structure of the present Nashville dome. The structural dome is growing as the basin enlarges and deepens. In our study the present structure of the Nashville dome disappears on all the no-load surfaces and cross-sections up to 150 minutes that were calculated.

A northward trending elongate dome once existed in the area of the Cumberland Plateau in Tennessee and it connected with the Jessamine dome of Kentucky. This dome is now isostatically suppressed. We suggest the association between this Cumberland Plateau dome and the Jessamine dome is controlled by basement structure related to the east continent gravity-magnetic high. It is possible that this basement non-isostatic uplift acted as a buttress to control the westward propagation of thrusting in Tennessee. The dome existed as early as Mississippian as evidenced by the coincidence between oil-bearing, mud mounds, and oolitic and porous zones on the north end of the Cumberland Plateau dome and oolitic bars on the southern end. Paleocurrent directions of Pennsylvanian conglomerates suggest that the dome was high enough to influence some aspects of sedimentation then. Although the Nashville dome is "merely" an isostatic uplift, other regional structures are not. The Jessamine dome does not disappear on no-load surfaces. The Clifton saddle-Pascola arch continues eastward to include the Hohenwald platform. The regional basins (Illinois, Appalachian, and Warrior) diminish in depth, but are present on no-load structure maps.

ACKNOWLEDGMENTS

We wish to thank G. Randy Keller for his constructive criticism of the manuscript. Donald R. Currey provided a new set of Lake Bonneville shoreline elevations for modeling the isostatic adjustment at Lake Bonneville.

REFERENCES CITED

Airy, G.B., 1855, On the computation of the effect of the attraction of the mountain-masses, as disturbing the apparent astronomical latitude of stations in geodetic surveys: Philos. Trans. R. Soc. London, v. 145, p. 101-104.

Bills, B.G., and May, G.M., 1987, Lake Bonneville: Constraints on lithospheric thickness and upper mantle viscosity from isostatic warping of Bonneville, Provo, and Gilbert Stage shorelines: Jour. Geophys. Res., v. 92, p. 11,493-11,508.

Brotchie, J.F., and Silvester, R., 1969, On crustal flexure: Jour. Geophys. Res., v. 74, p. 5240-5252.

Byers, C.W., 1977, Biofacies patterns in euxinic basins - a general model: Soc. Econ. Paleontologists and Mineralogists Spec. Pub. 25, p. 5-17.

Cathles, L.M., 1975, The Viscosity of the Earth's Mantle: Princeton Univ. Press, Princeton, 386 p.

Conant, L.C. and Swanson, V.E., 1961, Chattanooga Shale and related rocks of central Tennessee and nearby areas: U.S. Geol. Sur. prof. paper 357, 91 p.

Crittenden, M.D., Jr., 1963a, Effective viscosity of the Earth derived from isostatic loading of Pleistocene Lake Bonneville: J. Geophys. Res., v. 68, p. 5517-5530.

Crittenden, M.D., Jr., 1963b, New data on the isostatic deformation of Lake Bonneville: U.S. Geol. Sur., Prof. Paper 454-E, 31p.

Currey, D.R., 1982, Lake Bonneville: Selected features of relevance to neotectonic analysis: U.S. Geol. Sur. Open File Rpt., 82-1070, 31 p.

Currey, D.R., and Burr, T.N., 1988, Linear model of threshold controlled shorelines of Lake Bonneville: Utah Geol. and Miner. Survey, Misc. pub. 88-1, p. 104-110.

De Rito, R.F., Cozzarelli, F.A., and Hodge, D.S., 1986, A forward approach to the problem of nonlinear viscoelasticity and the thickness of the mechanical lithosphere: J. Geophys. Res. v. 91, p. 8295-8313.

Diment, W.H., and Urban, T.C., 1981, Average elevation map of the conterminous United States (Gilluly Averaging Method): U.S. Geol. Sur., Geophysical Invest. Map MAP GP-993, Scale 1/2,500,000.

Dorman, L.M., and Lewis, B.T.P, 1970a, Experimental isostasy 1. Theory of the determination of the earth's isostatic response to a concentrated load: J. Geophys. Research, v. 75, p. 3357-3365.

Dorman, L.M., and Lewis, B.T.P., 1970b, Experimental isostasy 2. An isostatic model for the U.S.A. derived from gravity and topographic data: J. Geophys. Research, v. 75, p. 3367-3386.

Fulton, L.P., 1979, Structure and isopach map of the New Albany-Chattanooga-Ohio Shale (Devonian and Mississippian) in Kentucky; Eastern Sheet: Kentucky Geol. Sur., Series XI, 1979, Scale 1/250,000.

Gilluly, J., Waters, A.C.,and Woodford, A.O., 1968, Principles of Geology: 3rd. Edition: W.H. Freeman and Co., San Francisco, 689 p.

Grohskopf, J.G., 1955, Subsurface geology of the Mississippi Embayment of southeast Missouri: Missouri Geological Survey Bulletin, 2nd series, v. 37, 133 p.

Gutenberg, B., 1941, Changes in sea level, postglacial uplift, and mobility of the Earth's interior: Bull. Geol. Soc. Am., v. 52, p. 721-772.

Handford, C.R., 1978, Monteagle Limestone (Upper Mississippian)-- Oolitic tidal-bar sedimentation in Southern Cumberland Plateau: AAPG Bull., v. 62, p. 644-656.

Hayford, J.F., and Bowie, W., 1912, The effect of topography and isostatic compensation upon the intensity of gravity: U.S. Coast Geod. Sur., Publ. 10, 132 p.

Heiskanen, W.A., and Vening Meinesz, F.A., 1958, The Earth and Its Gravity Field: McGraw-Hill, New York, 470 p.

Jackson, S.R., 1984, Depositional environment of the Lower Pennsylvanian Rockcastle Conglomerate, northern Cumberland Plateau, Tennessee: Unpubl. Masters Thesis, Vanderbilt Univ., Nashville, 148 p.

Keller, G.R., Bland, A.E., and Greenberg, J.K., 1982, Evidence for a major late Precambrian tectonic event (rifting?) in the eastern midcontinent region, United States: Tectonics, v. 1, p. 213-223.

Keller, G.R., Russell, D.R., Hinze, W.J., Reed, J.E., and Geraci, P.J., 1980, Bouguer gravity anomaly map of the East-Central Midcontinent of the United States: U.S. Nuclear Regulatory Commission, NUREG/CR-1663, 12 p.

Lumsden, D.N., Norman, C.D., and Reid, B.J., 1983, The Monteagle Limestone (Mississippian) in northcentral Tennessee: Petrology, porosity, and subsurface geology: Southeastern Geology, v. 24, p. 39-50.

McConnell, R.K., Jr., 1968, Viscosity of the mantle from relaxation time spectra of isostatic adjustment: J. Geophys. Res., v. 73, p. 7089- 7105.

Mc Dowell, R.C., Grabowski, G.J., Jr., and Moore, S.L., 1981, Geologic Map of Kentucky: The Eleventh Kentucky Geologic Survey, 4 sheets, scale 1:250,000.

Niskanen, E., 1939, On the upheaval of land in Fennoscandia: Ann. Acad. Sci. Fennicae, A III, v. 53, p. 1-30.

Nunn, J.A., Sleep, N.H., and Moore, W.E., 1984, Thermal subsidence and generation of hydrocarbons in Michigan Basin: AAPG Bull., v. 68, p. 296-315.

Peltier, W.R., 1984, The thickness of the continental lithosphere: Journal

Geophysical Research, v. 89, p. 11,303-11,316.
Potter, P.E., 1978, Structure and isopach map of the New Albany-Chattanooga-Ohio Shale (Devonian and Mississippian) in Kentucky: Central Sheet: Kentucky Geological Survey, Series X, 1978, Scale 1/250,000.
Potter, P.E., Wilson, E.N., and Zafar, J.S., 1984, Structure and thickness of the Devonian-Mississippian shale sequence in and near the Middlesboro syncline in parts of Kentucky, Tennessee, and Virginia: Kentucky Geological Survey, Report of Invest. 2, Series XI, 15p.
Pratt, J.H., 1855, On the attraction of the Himalaya Mountains, and of elevated regions beyond them upon the plumb-line in India: Philos. Trans. R. Soc. London, v. 145, p. 53-100.
Quinlan, G.M., and Beaumont, C., 1984, Appalachian thrusting, lithospheric flecture, and the Paleozoic stratigraphy of the Eastern Interior of North America: Canadian Journal of Earth Science, v. 21, p. 973-996.
Reesman, A.L., and Godfrey, A.E., 1981, Development of the Central Basin of Tennessee by chemical denudation: Zeitschrift fur Geomorphologie, v. 25, p. 437-456.
Reesman, A.L., Currey, D.R., and Bills, B.G., in manuscript, Regression analysis: A non-assumptive approach to the study of isostatic rebound at Lake Bonneville:
Rheams, K.F., 1981, Structural contour map showing configuration of the top of the Chattanooga Shale: Geol. Sur. of Alabama.
Rheams, K.F., and Neathery, T.L., 1984, Characterization and geochemistry of Devonian oil shale: North Alabama, North-west Georgia, and south- central Tennessee (A resource evaluation) Final Report; U.S. Dept. of Energy, Morgantown Energy Technology Center, National Technical Information Service, U.S. Dept. Commerce, Springfield, Va., 283 p.
Rice, C.L., 1984, Sandstone units of the Lee Formation and related strata in Eastern Kentucky: U.S. Geological Survey Prof. Paper 1151-G, 53 p.
Rich, J.L., 1951, Probable fundo origin of Marcellus-Ohio-New Albany-Chattanooga bituminous shales: Am. Assoc. Petroleum Geologists Bull., v. 35, p. 2017-2040.
Roedder, E., 1971, Fluid inclusion evidence on the environment of formation of mineral deposits of the southern Appalachian Valley: Economic Geology, v. 66, p. 777-791.
Schwalb, H.R., and Potter, P.E., 1978, Structure and isopach map of the New-Albany-Chattanooga-Ohio Shale (Devonian and Mississippian) in Kentucky: Western Sheet; Kentucky Geological Survey, Series X, 1978, Scale 1/250,000.
Simpson, R.W., Jachens, R.C., and Blakely, R.J., 1986, A new isostatic residual gravity map of the conterminous United States with a discussion on the significance of isostatic residual anomalies: Jour. Geophysical Research, v. 91, p. 8348-8372.
Stearns, R.G., and Reesman, A.L., 1986, Cambrian to Holocene structural and burial history of Nashville Dome: AAPG Bull., v. 70, p. 143-154.
Watts, A.B., and Ryan, W.B.F., 1976, Flexure of the lithosphere and continental margin basins: Tectonophysics, v. 36, p. 25-44.
Walcott, R.I., 1970, Isostatic response to loading of the crust in Canada: Can. J. Earth Sci., v. 7, p. 716-727.
Wanless, H.R., 1975, Illinois basin region, in McKee, E.D. and E.J. Crosby

coordinators, Introduction and regional analyses of the Pennsylvanian System, Part 1, of Paleotectonic investigations of the Pennsylvanian System in the United States: U.S. Geological Survey Professional Paper 853, pt. 1, 69-95.

Wilson, C.W., Jr, 1939, Probable connection of the Nashville and Ozark domes by a complementary arch: Jour. Geology, v. 47, p. 583-597.

Wilson, C.W., Jr., and R.G. Stearns, 1963, Quantitative analysis of Ordovician and younger structural development of Nashville Dome Tennessee: AAPG Bull, v. 47, p. 823-832.

Woollard, G.P., 1969, Regional variations in gravity: in The Earth's Crust and Upper Mantle: Pembrok Hart, Editor, Geophysical Monograph 13, Am. Geophysical Union, Washington, p. 320-341.

Zietz, I., and Bond, K., 1984, Aeromagnetic map of Tennessee and Kentucky: in color: U.S. Geological Survey Geophysical Invest. Map, Map-GP-959, scale 1:1,000,000.

ECHINOID BIOFACIES AND LITHOFACIES DISTRIBUTIONS IN THE UPPER EOCENE OF THE DOUGHERTY PLAIN, SOUTHWESTERN GEORGIA

BURCHARD D. CARTER

Department of Geology and Physics
Georgia Southwestern College
Americus, GA 31709

ABSTRACT

Paleoecologic and lithologic/petrologic investigation suggests that the late Eocene (Jacksonian) strata of southwestern Georgia were deposited on a shelf with changing facies distributions through time. In the early Jacksonian, carbonate and terrigenous sand predominated over the entire shelf. By the middle Jacksonian terrigenous deposition had ceased and deepening water and/or expanding grass banks caused the carbonate facies to become muddier over the central part of the shelf. In the region immediately northwest of the shelf break, at the edge of the Suwannee Strait, higher energy facies persisted through the entire Jacksonian. This may reflect continuing high current energy within the Strait, or simply propinquity to deep-water waves generated in the Strait and greater tidal influence. Higher energy persisted in a more nearshore zone as well, at least through deposition of the middle Jacksonian. The equivalent Ocala Group of peninsular Florida has also been reported to have experienced an expansion of muddy carbonate facies through the Jacksonian, but they were not prevalent until later (*Oligopygus wetherbyi* zone) than in Georgia (*O. haldemani* zone).

INTRODUCTION

The late Eocene (Priabonian or Jacksonian) carbonate rocks of the Dougherty Plain (Figure 1) in southwestern Georgia have received no systematic treatment since the general stratigraphic work of Cooke (1943), though J.G. Carter (1984) and Huddlestun and Hetrick (1986) summarized subsequent pertinent work. In addition, there has been little or no petrologic or sedimentologic work on these rocks other than minor thin section descriptions used to support a discussion of the substrate preferences of the echinoids (Carter and others, 1989). The present paper is a preliminary step toward characterizing the facies distributions of these limestones. Primary data are the ecological paleobiogeographical distributions of echinoid species supported by lithofacies and petrologic data.

A three-fold biostratigraphic subdivision of the Ocala Limestone in Florida (Hunter, 1976; McKinney and Jones, 1983; Croft and Shaak, 1985; McKinney and Zachos, 1986) can be extended northward and applied in Georgia (Carter and Hammack, 1989). However, it should be noted that the *Oligopygus* species used effectively as guide fossils in Florida are not common in Georgia. McKinney (1984; 1985) has argued convincingly that the *Oligopygus* species used in the Florida biozonation represent an evolutionary lineage, and thus their zones are temporally and not facies controlled. In Georgia the *Oligopygus phelani* zone (zone 1 herein) approximately corresponds to the Clinchfield Sand and perhaps

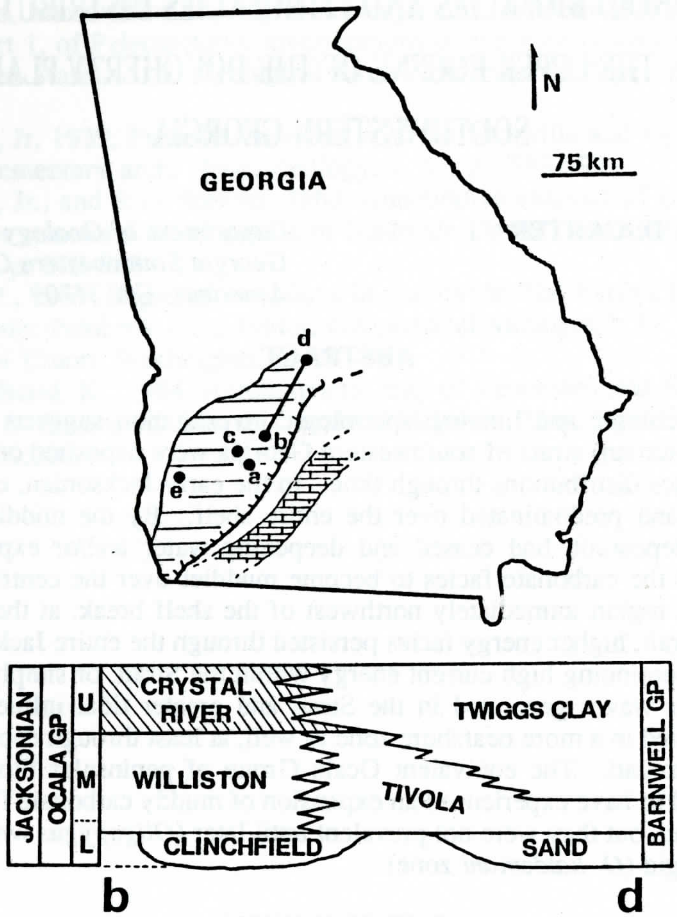


Figure 1. Locality map and cross-section of late Eocene (Priabonian/Jacksonian) strata of southwestern Georgia (modified from Huddlestun and Hetrick, 1986). Hatching of Crystal River on the cross-section reflects poor exposure. Localities explained in Table 1. The hatched region on the map is the Dougherty Plain, which is the region of southwestern Georgia where Eocene rocks are exposed. The fine brick pattern on all maps herein represents the Suwannee Strait.

part of the Williston and Tivola Limestones. Carter and Hammack (1989) informally refer to this as the lower Jacksonian. The *O. haldemani* zone (zone 2 herein) is equivalent to the remainder of the Williston and Tivola Limestones (middle Jacksonian of Carter and Hammack). The *O. wetherbyi* zone (upper Jacksonian of Carter and Hammack; zone 3 herein) roughly equates with the Crystal River Formation. The latter is observed primarily as residual chert boulders in the Dougherty Plain (Figure 1), which is an active karst region. Echinoids of the late Jacksonian of Florida, including *O. wetherbyi*, are typically found only in such boulders in this part of Georgia. Based upon autecological and stratigraphic analyses of echinoid species, Carter and Hammack (1989), Carter and others (1989), and Carter (1987a) suggested that the distinctiveness of the echinoid faunas in Georgia and central Florida results from: 1) missing strata in Georgia (the *O. wetherbyi* equivalents) and 2) substrate differences (muddier conditions in Georgia earlier than in Florida). In central Florida, several theses (Fenk, 1979;

Sharpe, 1980; Zachos, 1978) and published work (Randazzo and Saroop, 1976) have characterized the Ocala petrologically. They report a fining-upward sequence with carbonate sand dominating the *O. phelani* and *O. haldemani* zones and muddy sediments dominating the *O. wetherbyi* zone. Carter and others (1989) characterized the echinoids of the *O. haldemani* zone equivalent in Georgia as being dominated by mud-tolerant species. The lithofacies and petrologic work on the Jacksonian strata of southwestern Georgia presented below also suggests the prevalence of mud-rich substrata during that time.

Table 1 lists and describes the localities mentioned in the text, figures, and later tables.

Table 1. Descriptions of localities mentioned in the text, figures, and other tables.

Locality Symbol, Maps/Tables	USGS 7.5' Topo. Map	Lat./Long.	Locality Description
a	Albany West	31° 36'/84° 08'	Power plant on N edge of Albany, Dougherty Co.
a1	Albany West	-----	Water wells near Albany, Dougherty Co.
b	Neyami	31° 46'/84° 09'	Leesburg Quarry ~3 mi. NNE of Leesburg at Starksville, Lee Co.
b1	Leesburg	-----	Water wells near Leesburg, Lee Co.
c	Sasser	31° 43'/84° 15'	Oakland Quarry ~8 mi. NW of Albany in Lee Co.
d	Perry East	32° 24'/83° 38'	Medusa Cement Co. Quarry at Clinchfield, Houston Co.
e	Bancroft	31° 27'/84° 47'	Arlington Quarry on Spring Creek, Calhoun Co.
f	?Doerun	?	Test well Core, GGS 3545 near Moultrie, Colquitt Co.
g	Baconton North	~31° 23'/84° 10'	Water well N of Baconton, Mitchell Co.
h	Saffold	31° 07'/85° 03'	Railroad bridge over Chattahoochee River at Saffold, Early Co.
i	Bainbridge	-----	Localities along Flint River N and S of Bainbridge, Decatur Co.
j	?	?	Ichawaynochaway Creek, Baker Co.
l	?Hopeful	?	Normans Ferry/Dry Bread Shoal on Flint River, Mitchell/Baker Co.
m	Cobb	31° 58'/83° 55'	Flint River railroad bridge W of Coney, Crisp Co.
n	Albany East	~31° 37'/84° 03'	Flint River in NE corner of Dougherty Co.
o	?Colquitt	?	Spring Creek near Colquitt, Miller Co.

ECHINOID PALEOECOLOGY

Echinoid species often have specific, easily determined sediment preferences, and the geographic distributions of late Eocene echinoid assemblages agree well with the distributions of mud-rich and clean sand sediments discussed in the lithofacies section below. Figure 2 is a summary of the substrate preferences of the Jacksonian irregular echinoids of southwestern Georgia, modified from Carter and others (1989). Their interpretations of the echinoids' association with certain sediment types was based upon test characteristics (principally the longitudinal profile of tests, aboral tubercle density, and morphology of lateral tubercles, fascioles, and petals), comparison with Recent relatives, stratigraphic position (particularly in Florida), and petrologic examination of the sediment within and adhering to the fossils. The sediment outside an echinoid's test clearly reflects the final depositional environment of the test, and is thus of direct interest to this study. That inside the test is a mixture of the gut content at death, the sediment at the site of death, and the sediment of the final resting place, assuming the latter two to have been different. In virtually all cases the sediment within and outside the tests was indistinguishable; the only notable exceptions were

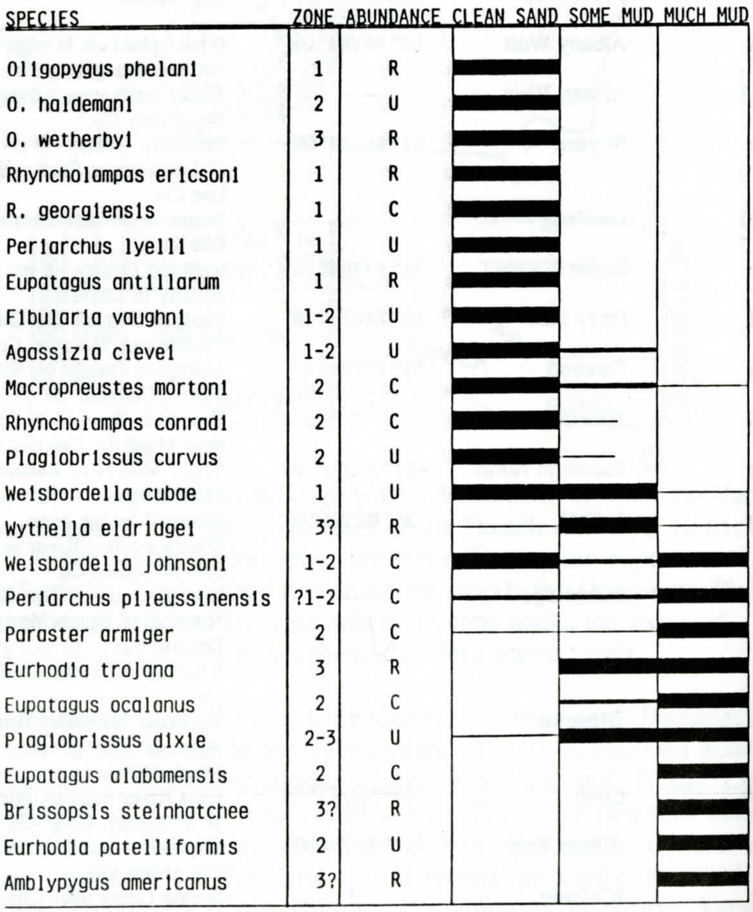


Figure 2. Substrate preferences of Jacksonian echinoids in Georgia. Data from Carter and others (1989). Abundance symbols refer to abundances in Georgia only — "C" is common; "U" is uncommon; "R" is rare. Thickness of lines represents commonness in each sediment type. "Clean Sand" is quartzarenite or biosparite with less than 10% of interstices containing mud; "some mud" implies 10-25% mud- filled interstices; "much mud" means more than 25% (often >50%) of interstices with mud. All samples examined were grain-supported sediment.

specimens in which geopetal structures clearly indicated transport. Both Croft and Shaak (1985) and Carter and others (1989) summarize abundant evidence that echinoid tests generally are not transported far from their environments of life when they die. Because these samples typically do not come from known stratigraphic horizons their utility is restricted to suggesting the types of sediment present at a locality at unspecified times during the accumulation of that sediment.

Notice that all species restricted to the lower Jacksonian except possibly *Weisbordella* spp. are sand-dwellers, and that none of the mud-tolerant species is known from the lower Ocala. In fact, the *Weisbordella* spp. from the lower Ocala are only tentatively assigned, and may represent an unnamed species of *Neolaganum*, other species of which were sand-dwellers. Middle Jacksonian and upper Jacksonian species are mixed sand- and mud-dwellers, with the latter more common. The fact that there were populations of echinoids inhabiting the sandy substrata throughout the middle and upper Jacksonian suggests that those sands were persistent facies rather than the result of short-term depositional events, such as storms, affecting otherwise quiet water.

Carter (1987b) summarized the geographic distribution of late Eocene echinoid species in the southeastern United States. Table 2 and Figure 3 represent a summary of the distribution of lower Jacksonian echinoids and their substrate preferences. Note that all localities have faunas of sand-dwelling echinoids only.

Table 3 and Figure 4 present the geographic distributions for middle Jacksonian species. The northernmost and the two westernmost localities, where sand-dwellers dominate, are Medusa Quarry, Saffold, and Arlington respectively (Figure 4 locations d,e, and h). The echinoid data indicate that these localities were dominated by sandy facies. Note also that those localities nearest the Suwannee Strait were inhabited primarily by sand-dwelling echinoids. Between the Strait and the present northwesternmost exposures is a region of primarily mud-tolerant assemblages, with some species which preferred sand.

Table 4 and Figure 5 summarize the few occurrences of upper Jacksonian echinoids in Georgia. The outcrops are typically chert boulders in residuum, but the southernmost ones were probably in situ limestone outcrops visited by C.W.

Table 2. Echinoid geographic distribution for the *Oligopygus phelani* zone equivalent (zone 1). Combining these data with the substrate preferences in Figure 2 allows construction of Figure 3. Localities are explained in Table 1.

a1 - *Periarchus pileussinensis, Weisbordella cubae, Fibularia vaughni*
b1 - *Periarchus pileussinensis, Weisbordella johnsoni*[1], *Weisbordella cubae, Rhyncholampas ericsoni, Rhyncholampas georgiensis*
d - *Periarchus pileussinensis, Weisbordella johnsoni*[1], *Eupatagus antillarum*?
e - *Periarchus lyelli, Rhyncholampas georgiensis*
g - *Oligopygus phelani, Weisbordella johnsoni*[1]
h - *Rhyncholampas georgiensis, Weisbordella johnsoni*[1,2], *Agassizia clevei*[2], *Fibularia vaughni*[2]
i - *Rhyncholampas georgiensis, Agassizia clevei*[2], *Fibularia vaughni*[2], *Weisbordella johnsoni*[1,2]

---

1 - Morphologically distinct from typical *W. johnsoni*, and possibly an undescribed *Neolaganum*.
2 - Uncertain stratigraphic position; possibly belong in the middle Ocala.

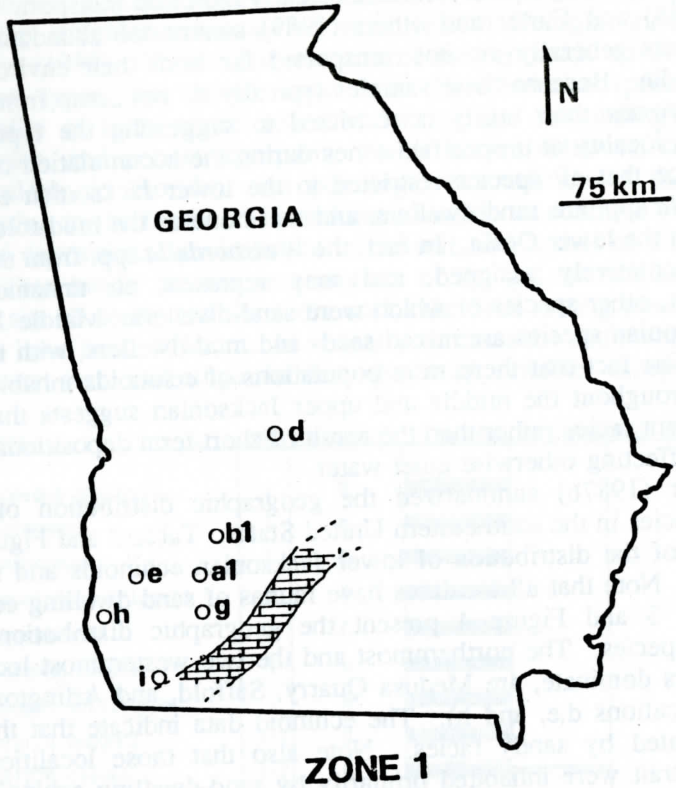


Figure 3. Geographic distribution of echinoid assemblages with specific substrate preferences, lower Jacksonian (zone 1). Open circles symbolize assemblages dominated by sand-dwellers. Locality data for echinoid species from Carter (1987b) and references therein. Substrate preferences from Carter and others (1989). Locality symbols explained in Table 1.

Cooke (1959). These localities are now apparently flooded by Lake Seminole. Though the number of samples is small and their geographic distribution not ideal, it appears from Figure 5 that during deposition of upper Jacksonian sediment, carbonate sand may have been prevalent along the northwest flank of the Suwannee Strait, with muddier sediment farther to the northwest. Specimens of *Oligopygus wetherbyi*, obviously transported because they are not in life-position, have been identified from cores in the northern edge of the Strait. This also suggests that sandy conditions prevailed upslope. This pattern is the same as that suggested for the middle Jacksonian.

LITHOFACIES AND PETROLOGY

Data

Lithologic and petrologic characterization of various localities exposing late Eocene rocks allows assessment of the paleogeographic distribution of facies, particularly the distribution of mud-rich versus sandy carbonate sediments. Petrographic information comes primarily from three sources: hand samples, thin sections from measured sections, and thin sections of echinoids (often of uncertain

Table 3. Echinoid geographic distribution for the *Oligopygus haldemani* Zone equivalent (zone 2). Combining these data with the substrate preferences in Figure 2 allows construction of Figure 4. Localities are explained in Table 1.

a - *Paraster armiger, Eurhodia patelliformis, Eupatagus alabamensis, Weisbordella johnsoni, Plagiobrissus dixie, Rhyncholampas conradi, Plagiobrissus curvus, Macropneustes mortoni*

b - *Periarchus pileussinensis, Weisbordella johnsoni, Paraster armiger, Eurhodia patelliformis, Eupatagus alabamensis, E. ocalanus, Macropneustes mortoni, Rhyncholampas conradi, Plagiobrissus curvus*? (Rare), *Oligopygus haldemani* (Rare)

d - *Periarchus pileussinensis, Oligopygus haldemani, Macropneustes mortoni, Paraster armiger*

e - *Oligopygus haldemani, Agassizia clevei, Plagiobrissus curvus*?

h - *Oligopygus haldemani*

i - *Oligopygus haldemani, Agassizia clevei*[1], *Plagiobrissus curvus, Rhyncholampas conradi*[1], *Fibularia vaughni*[1], *Weisbordella johnsoni, Eupatagus ocalanus*

j - *Eurhodia patelliformis*

l - *Oligopygus haldemani, Weisbordella johnsoni, Eurhodia patelliformis*

m- *Eupatagus alabamensis, Plagiobrissus dixie, Weisbordella johnsoni, Eurhodia patelliformis*

n - *Macropneustes mortoni, Paraster armiger, Weisbordella cubae*

o - *Weisbordella johnsoni, Plagiobrissus dixie, Rhyncholampas conradi*[1]

---

[1] - Uncertain stratigraphic position; possibly belong in lower or upper Ocala.

stratigraphic position) from various localities. I have not yet measured and sampled most exposures of the Jacksonian strata in the region, and rely herein on hand-sample descriptions. For the Tivola Limestone this is probably adequate, as it is a very distinctive lithology (ectoproct calcarenite). Similarly, the Clinchfield Sand in the Dougherty Plain is invariably a fossiliferous quartzarenite; the only thin section examined adds little to this information except that a few peloids and fragments of Foraminifera, ectoprocts, scallops, and other molluscs are scattered within it. However for the Williston Formation (and its informally defined Muckalee member) several students and I have measured sections and sampled them for thin section analysis (Figure 6 and Appendices 1 and 2). Only five or six samples have been taken from each of three localities because the exposed strata are so thin. The third source of petrographic data is part of the set of thin sections used by Carter and others (1989) in their study of echinoid paleoecology. Some 75 of their sections are from these three and other late Eocene exposures in southwest Georgia.

A quarry northeast of Leesburg, Lee County, Georgia (locality b in Figure 1, section b on Figure 6, and Appendix 1A) is the proposed type section of the informally named Muckalee member of the Williston Formation (Huddlestun, 1981). Slightly less than 5m of thick-bedded, soft, white limestone are exposed. Figure 6b is based upon field examination, and the size of the brick pattern is proportional to the field description of the coarseness of the limestone. In general, thin sections (Appendix 2) from the units described in the field as "coarse" are biopelsparites with only small amounts of carbonate mud. Those from "fine" units are typically biopelmicrites and very poorly washed biopelsparites. There is

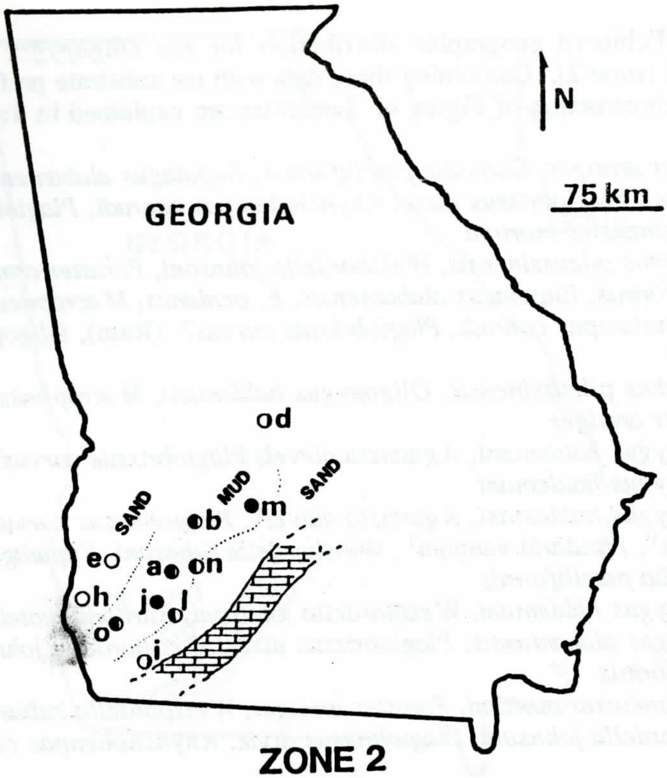


Figure 4. Geographic distribution of echinoid assemblages with specific substrate preferences, middle Jacksonian (zone 2). Open circles symbolize assemblages dominated by sand-dwellers. Filled circles symbolize assemblages dominated by mud-tolerant species. Amount of fill is proportional to the fraction of species preferring muddy sediment. Locality and substrate preference data as in Figure 3. Localities explained in Table 1. Dotted lines separate sand-bottom communities from mud-tolerant communities.

Table 4. Echinoid geographic distribution for the *Oligopygus wetherbyi* zone equivalent (zone 3). Combining these data with the substrate preferences in Figure 2 allows construction of Figure 5. Localities are explained in Table 1.

e - *Eurhodia trojana, Schizaster*? sp. (usually mud-dwellers)
h - *Amblypygus americanus*
i - *Oligopygus wetherbyi, Rhyncholampas conradi*
j - *Oligopygus wetherbyi, Rhyncholampas conradi*[1]
l - *Wythella eldridgei*
m- *Eurhodia trojana*
o - *Eurhodia trojana, Plagiobrissus dixie, Rhyncholampas conradi*[1]

---

1 - Uncertain stratigraphic position; possibly belong in the middle Ocala.

little difference in average grain size of the sand component of these rocks except that skeletons in the mud-rich sediments are more often whole, and therefore coarser. No matrix-supported sediments have been observed here, even among

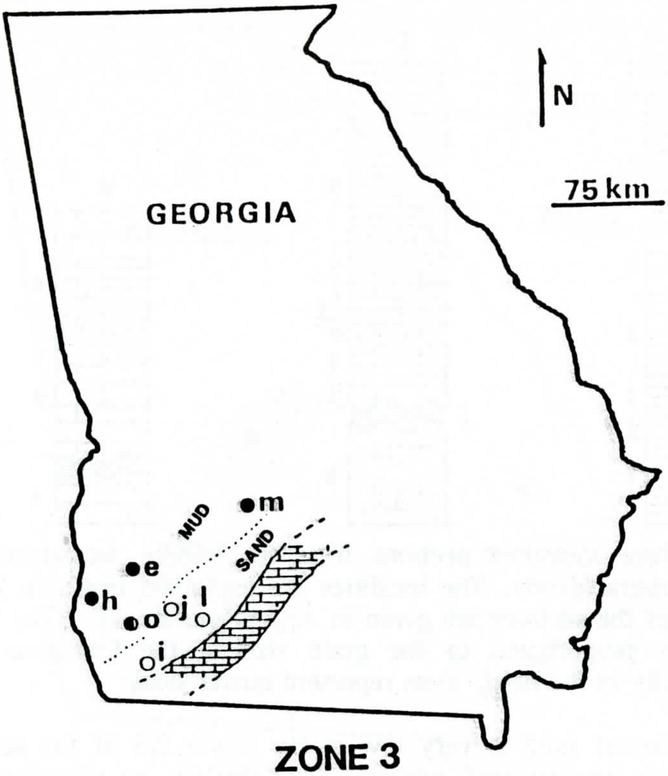


Figure 5. Geographic distribution of echinoid assemblages with specific substrate preferences, upper Jacksonian (zone 3). Symbols as in Figure 4. Localities explained in Table 1.

the 50 echinoid thin sections studied. As can be inferred from the brick pattern on Figure 6b, the sediments are predominantly rich in carbonate mud, suggesting low energy conditions. The interbedded cleaner biosparite units indicate locally higher energy. There is apparently a trend for the carbonate sand units to be more common in the lower part of the exposure.

Some 25km southwestward, in the southwestern corner of Lee County (Figure 1c, Figure 6c, and Appendix 1B) is a quarry at Oakland, Georgia operated by the Martin-Marietta Company. During mining operations Clinchfield Sand boulders up to 2m in diameter were dredged from below the water table. The walls presently expose about 3.5m of the Muckalee member of the Williston. An additional 2-3m (approximately) of this unit was previously exposed, but the mining company has since quit pumping water and this is now flooded. The Muckalee here (thin sections of 5 samples - Appendix 2) is virtually all biopelmicrite (sometimes matrix supported) or very poorly washed biopelsparite. Only the lowermost 20cm measured was fairly clean biosparite. Thus energy was apparently consistently lower here than at the Leesburg Quarry unless the lower covered interval is primarily biosparite.

Another 50km southwestward is an abandoned quarry near Arlington, Calhoun County, Georgia (Figures 1e and 6e, and Appendix 1C). Six thin sections were made from samples of these rocks (Appendix 2). The Williston Limestone is exposed here, and is primarily comprised of biopelsparites. The amount of

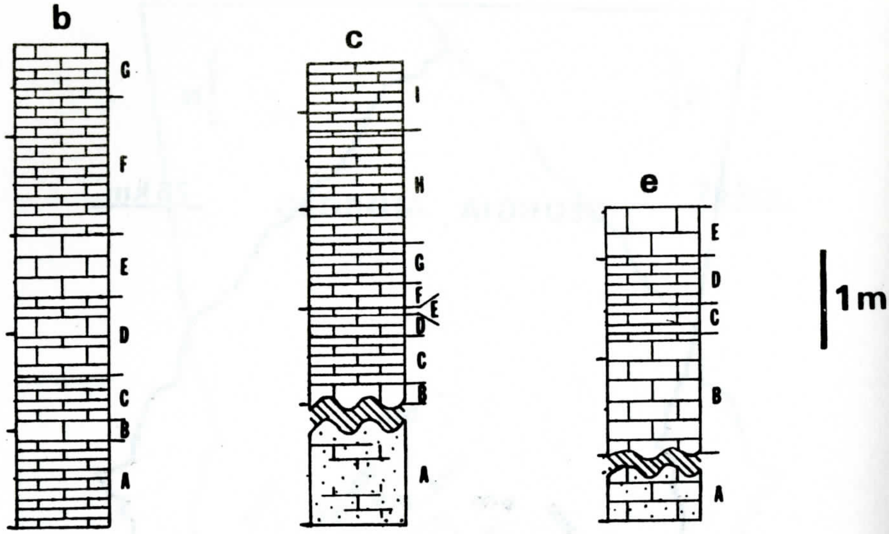


Figure 6. Three measured sections from the middle Jacksonian which were analyzed petrographically. The localities are explained in Table 1 and detailed descriptions of the sections are given in Appendices 1A-1C. The "thickness" of the bricks is proportional to the grain size of the limestone as estimated macroscopically in the field. Dots represent quartz sand.

admixed carbonate mud is very low in the lower 2/3 of the section, but the uppermost beds are less well washed. Nevertheless, no allomicrites have been found, only poorly washed biopelsparites. The measured section is only slightly over 2.5m thick, and is underlain by an unknown thickness of apparently similar lithology, now covered with water. The lowest beds mined are also flooded, but dredged boulders can be identified from the fossils they contain — principally *Periarchus lyelli.* These contain coarse quartz sand which is reminiscent of the Clinchfield.

Interpretation

The sequence of lithologies in the late Eocene suggests gradually decreasing energy from the Clinchfield Sand and Tivola Limestone calcarenite through the lower sandier(?) part of the Williston, to the muddier upper part of the Williston and the Muckalee member. Figure 7 is a facies map of southwest Georgia for the time during the deposition of lower Jacksonian (zone 1) sediments. Coarse calcarenite and quartzarenite are virtually ubiquitous where thin sections and samples have been examined (locations a, b, e, and d, among others), as would be predicted from Figure 3. Location g, a well near Baconton, Mitchell County, lacks quartz sand, but is interpreted as either carbonate or terrigenous sand on paleoecologic evidence. Location f (a well near Moultrie, Colquitt County) penetrates the Suwannee Strait. The Suwannee Strait was a paleobathymetric low where pelagic carbonate mud was deposited. Eocene rocks in the Strait include a dominantly planktonic biota with only rare, transported macrofossils (pers. obs. and Paul Huddlestun, pers. comm. 1984).

Workers at least as early as Cooke (1943) have recognized that during at least part of the Jacksonian the shoreline was well north of the exposures studied

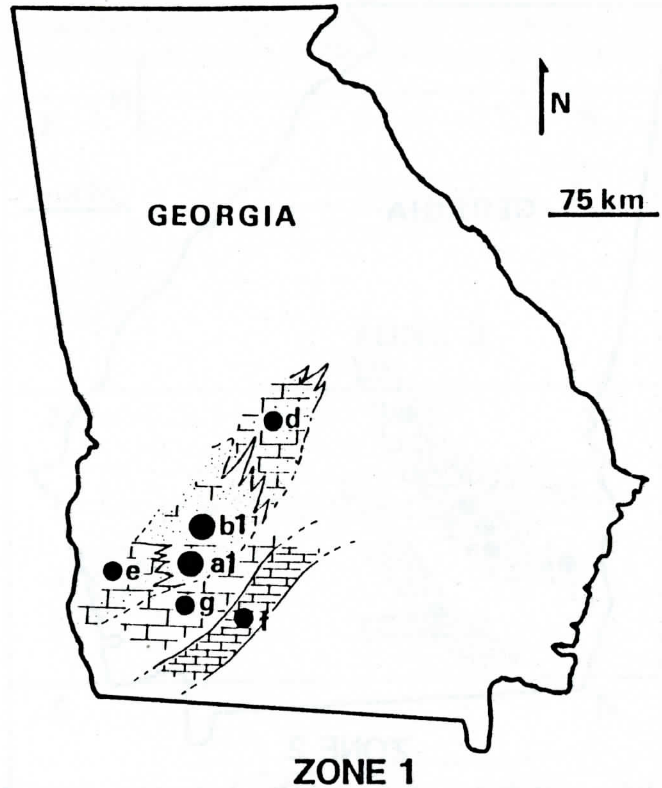


Figure 7. Localities and facies of the lower Jacksonian (zone 1). Symbols as in Figure 6. Localities explained in Table 1.

herein (at least 75km north of location d on Figure 7). Therefore it is not clear whether the quartz sand represents an initial transgressive facies, reworking of the underlying Claibornian sands, or sediment derived from the Piedmont during the Jacksonian and introduced far from the shore.

Figure 8 represents facies during deposition of middle Jacksonian (zone 2) rocks. Fairly coarse calcarenite persists in the northeastern and southwestern parts of the study area (Medusa Quarry, Houston County — location d, and Arlington, Calhoun County — location e) coinciding with the western area of sand-dwelling echinoid assemblages (Figure 4). Pelagic sediment continued to accumulate in the Suwannee Strait (location f). Over a broad central area, however, quieter water allowed carbonate mud to accumulate along with skeletal debris and peloids. Interspersed with these muddy bottoms were cleaner carbonate sands. These lithofacies were the habitats for the mixed mud-tolerant and sand-dwelling echinoid communities (Figure 4). As discussed in the section on paleoecology, the higher energy responsible for these sands was apparently not simply a result of storms or some other short-lived process, but of more persistent conditions. Perhaps they represent unbound sediment between muddy grass banks, as occurs presently in the nearshore part of Hawk Channel in the Florida Keys.

Although uppermost Jacksonian beds are not exposed in the region, the paleoecological evidence indicates continued low-energy, muddy conditions toward the center of the shelf, and sandier facies at the edge of the Suwannee

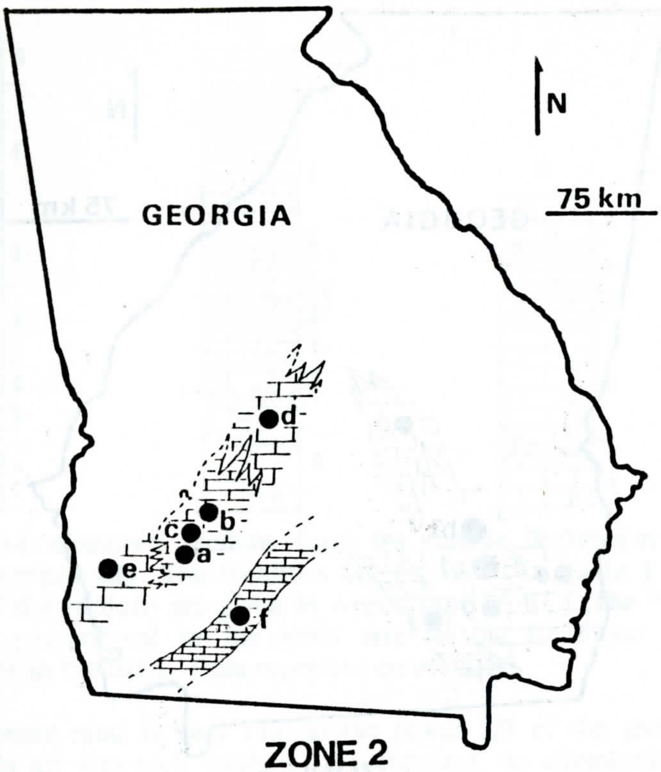


Figure 8. Localities and facies of the middle Jacksonian (zone 2). Symbols as in Figure 6. Localities explained in Table 1.

Strait. Croft and Shaak (1985) suggested the possibility of either increasing prevalence of grass banks or of transgression controlling the increased carbonate mud in the upper Jacksonian of Florida. From this study alone I cannot choose between these alternatives. However, carbonate sedimentologists in Florida (Fenk, 1979; Randazzo and Saroop, 1976; Sharpe, 1980; Zachos, 1978) have preferred transgression, and published sea-level curves (Vail and others, 1977; Haq and others, 1987) indicate that there was a global sea-level rise during the late Eocene. There is, of course, no reason that increasing depth and expanding grass banks should be mutually exclusive events. In Florida, dominantly muddy conditions were not established until deposition of the *O. wetherbyi* zone. This was later than in Georgia where mud-rich sediment was already accumulating during deposition of the *O. haldemani* zone.

Taken together, the paleoecologic and lithologic data can be used to reconstruct the gross facies distributions in southwestern Georgia in the late Eocene. At least after muddy sediment began accumulating on the shelf there was apparently a bank-type, rather than ramp-type facies distribution (Wilson, 1975). A distinct slope break at the edge of the strait is indicated by isopach and structure contour maps of Paleogene sediments (Herrick and Vorhis, 1963; Aurora, 1984). Figures 4 (locations i and l) and 5 (locations i,j, and l) indicate a carbonate sand facies at the shelf edge (northwest flank of the Suwannee Strait), finer pelagic sediment "basinward" in the Strait, and fine sediment also in the shallow lagoon landward of the shelf edge, a situation analogous to the present western margin of the Bahamas. However, this lagoon became sandier again farther landward, at least during middle Jacksonian deposition (Figure 4, locations e and h). Figure 9 summarizes the depositional history of the Jacksonian in Georgia.

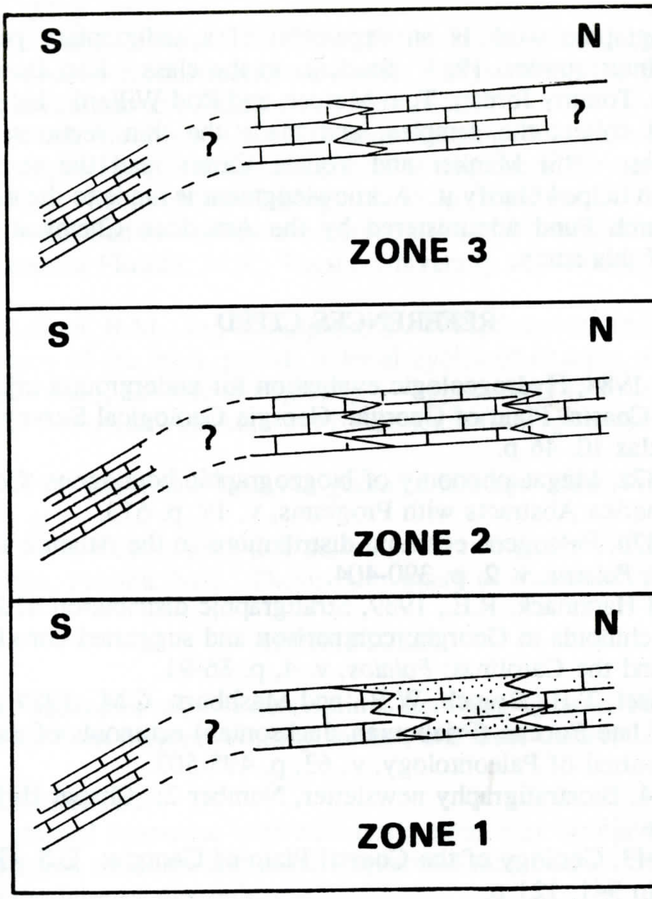


Figure 9. Schematic cross sections from the Suwannee Strait (in the south) north-northwestward across the Jacksonian shelf. Thick brick pattern implies sandy sediment, thin brick pattern implies muddier sediment. Not to scale.

CONCLUSIONS

1) Petrographic and paleoecologic investigations both suggest that the Jacksonian sediments of southwestern Georgia were deposited on a shelf, the central part of which became progressively lower in energy through time.
2) Deepening water and/or expanding grass banks, with attendant current baffling and sediment binding, caused the facies to become muddier through time on the central shelf. This upward increase in muddy facies is found also in the Ocala Group of Florida, but mud-rich facies did not become established there until later.
3) In the region immediately northwest of the shelf break, at the edge of the Suwannee Strait, higher energy facies persisted throughout Jacksonian deposition. This may reflect high current energy within the Strait, or stronger waves and tidal currents at its edge.
4) Higher energy persisted in a more nearshore zone as well, at least through the

deposition of middle Jacksonian sediments.

ACKNOWLEDGMENTS

The petrographic work is an expansion of a sedimentary petrology lab exercise from winter quarter, 1987. Students in the class - Ken Downing, Scott Drew, Jim Fineis, Tommy Jordan, Tom Mercer, and Rod Willard - helped measure the sections and collect the samples, and made the thin sections from those measured sections. Phil Manker and Yonnie Carter read the text and made suggestions which helped clarify it. Acknowledgment is made to the donors of the Petroleum Research Fund administered by the American Chemical Society for partial support of this study.

REFERENCES CITED

Aurora, R. (ed.), 1984, Hydrogeologic evaluation for underground injection control in the Coastal Plain of Georgia: Georgia Geological Survey Hydrologic Atlas 10, 46 p.

Carter, B.D., 1987a, Megataphonomy of biogeographic boundaries: Geological Society of America Abstracts with Programs, v. 19, p. 613.

Carter, B.D., 1987b, Paleogene echinoid distributions in the Atlantic and Gulf Coastal Plains: *Palaios*, v. 2, p. 390-404.

Carter, B.D., and Hammack, R.E., 1989, Stratigraphic distribution of Jacksonian (Priabonian) echinoids in Georgia: comparison and suggested correlations with Florida and the Carolinas: *Palaios*, v. 4, p. 86-91.

Carter, B.D., Beisel, T.H., Branch, W.B., and Mashburn, C.M., 1989, Substrate preferences of late Eocene (Priabonian, Jacksonian) echinoids of the eastern Gulf Coast: Journal of Paleontology, v. 63, p. 495-503.

Carter, J.G., 1984, Biostratigraphy newsletter, Number 2: Chapel Hill, North Carolina, 38 p.

Cooke, C.W., 1943, Geology of the Coastal Plain of Georgia: U.S. Geological Survey Bulletin 941, 121 p.

Cooke, C.W., 1959, Cenozoic echinoids of eastern United States: U.S. Geological Survey Professional Paper 321, 106 p.

Croft, M. and Shaak, G.D., 1985, Ecology and stratigraphy of the echinoids of the Ocala Limestone (late Eocene): Tulane Studies in Geology and Paleontology, v. 18, p. 127-143.

Fenk, E.M., 1979, Sedimentology and stratigraphy of middle and upper Eocene carbonate rocks, Lake, Hernando, and Levy Counties, Florida: M.S. Thesis, University of Florida, Gainesville, 133 p.

Haq, B.U., Hardenbol, J., and Vail, P.R., 1987, Chronology of fluctuating sea level since the Triassic: Science, v. 235, p. 1156-1167.

Herrick, S.M., and Vorhis, R.C., 1963, Subsurface geology of the Georgia Coastal Plain: Georgia Geological Survey Information Circular 25, 78 p.

Huddlestun, P.F., 1981, Correlation chart - Georgia Coastal Plain: Georgia Geological Survey Open-File Report 82-1.

Huddlestun, P.F. and Hetrick, J.H., 1986, Upper Eocene stratigraphy of central and eastern Georgia: Georgia Geological Survey Bulletin 95, 78 p.

Hunter, M.E., 1976, Biostratigraphy, *In* Hunter, M.E. (ed.), Tertiary carbonates, Citrus, Levy, Marion Counties, west central Florida: Southeastern

Geological Society Guidebook 18, p. 66-87.

McKinney, M.L., and Jones, D.S., 1983, Oligopygoid echinoids and the biostratigraphy of the Ocala Limestone of peninsular Florida: Southeastern Geology, v. 24, p. 21-29.

McKinney, M.L., and Zachos, L.G., 1986, Echinoids in biostratigraphy and paleoenvironmental reconstruction: a cluster analysis from the Eocene Gulf Coast (Ocala Limestone): *Palaios*, v. 1, p. 420-423.

Randazzo, A.F. and Saroop, H.C., 1976, Sedimentology and paleoecology of middle and upper Eocene carbonate shoreline sequences, Crystal River, Florida, U.S.A.: Sedimentary Geology, v. 15, p. 259-291.

Sharpe, C.L., 1980, Sedimentological interpretation of Tertiary carbonate rocks from west central Florida: M.S. Thesis, University of Florida, Gainesville, 170 p.

Vail, P.R., Mitchum, R.M., and Thompson, S., 1977, Seismic stratigraphy and global changes of sea level, part 4: Global cycles of changes of sea level, *In* Payton, C.E. (ed.), Seismic stratigraphy — applications to hydrocarbon exploration: American Association of Petroleum Geologists memoir 26, p. 83-93.

Wilson, J.L., 1975, Carbonate facies in geologic history: New York, Springer-Verlag, 471 p.

Zachos, L.G., 1978, Stratigraphy and petrology of two shallow wells, Citrus and Levy Counties, Florida: M.S. Thesis, University of Florida, Gainesville, 105 p.

APPENDICES

Appendix 1A. Measured section description, Leesburg Quarry (Figure 6, locality b). Measured by Tom Mercer, Tommy Jordan, and Burt Carter, January 29, 1988.

UNIT	DESCRIPTION
G	Fine grained limestone with thin terrigenous clay stringers and interbeds. Fossils less abundant and more fragmented than below. Sample LQ 4.5 is from the base of this unit. 4.4 to 4.9 m above base.
F	Fine grained limestone with terrigenous clay interbeds up to 20 cm. Fossils are generally whole. *Rhyncholampas conradi*, *Amusium ocalanum*, *Phyllacanthus mortoni* spines identified. Some *R. conradi* are not in life position. Sample 3.6. 3.0-4.4 m above base.
E	Coarse grained limestone. Fossils mostly fragmented except toward top, where they lie horizontally. Sample 2.8. 2.35-3.0 m.
D	Coarse, crystalline limestone. Two thin clay beds. Some large fossil fragments, particularly *Amusium ocalanum*, randomly oriented. ?*Rhyncholampas conradi*. Some thin beds may have a fine carbonate matrix. Covered from 2.2-2.3 m. Sample 1.8. 1.55-2.35 m.
C	Fine grained limestone with some interbeds of crystalline limestone. Fossil fragments mostly randomly oriented, but typically whole or large fragments. Bivalves including scallops, echinoid fragments including *Macropneustes mortoni*. Nodular character to bedding. Not sampled.

1.1-1.55 m.

B Coarse grained, crystalline, fossiliferous limestone. Fossil fragments randomly oriented. *Phyllacanthus mortoni*. Sample 0.9- 1.0.
0.9-1.1 m.

A Fine grained limestone with ectoprocts, ribbed scallops, and echinoid fragments including *Phyllacanthus mortoni.* Size and abundance of fossils increases toward top. Sample 0.4.
0-0.9 m.

Appendix 1B. Measured section description, Oakland Quarry (Figure 6, locality c). Measured by Rod Willard, Ken Downing, and Burt Carter, January 29, 1988.

UNIT DESCRIPTION

I Soft, fine grained limestone with scattered fossil fragments, both ribbed scallops and larger forams identifiable. Vertical and horizontal burrows obvious. Irregular bedding averaging 25 cm thick.
2.8-3.45 m.

H Hard, dense, fine grained limestone. Vertical and horizontal burrows obvious. Fossil fragments randomly oriented. Whole *Paraster armiger.* Sample OQ 4.
1.65-2.8 m.

G Fine grained, soft chalky burrowed limestone. Irregular bedding averages 15 cm. Fossils not obvious.
1.25-1.65 m.

F Hard, dense, fine grained limestone. Not obviously burrowed. One spatangoid, upside down.
1-1.25 m.

E Soft, chalky, fine grained limestone with vertical burrows. Scattered fossil fragments.
0.9-1 m.

D Somewhat harder fine grained limestone. Fossil fragments and whole fossils fairly abundant: scallops and other molluscs, large forams. Sample OQ 3.
0.7-0.9 m.

C Crumbly "nodular" limestone in softer, chalky matrix. All fine grained. Vertical burrows obvious. Sample OQ 2.
0.2-0.7 m.

B Hard, dense, coarse grained, crystalline limestone. Randomly oriented fossil fragments obvious including echinoids, scallops, and large forams. Sample OQ 1.
0-0.2 m.

INTERVAL > 2-3 m FLOODED.

A (Boulders dredged from below water level.) Calcareous quartz sandstone with abundant molluscs (including scallops) and the echinoids *Periarchus pileussinensis*, *Rhyncholampas georgiensis*, and ?*Weisbordella* sp. Sample OQ X.

Appendix 1C. Measured section description, Arlington Quarry. Measured by Jim Fineis, Scott Drew, and Burt Carter, February 12, 1988.

UNIT DESCRIPTION

E Very hard, dense, coarse, crystalline limestone. Larger forams, fragments of echinoderms about the only identifiable fossils. Sample ARL 3 at 2.4 m. and Sample ARL 5 at 2.5 m.
2-2.6 m.

D Hard, dense, interbedded fine and coarse grained limestone. Abundant large forams, ectoprocts, snails, and echinoids, including *Oligopygus haldemani* and tiny ?neolaganids. Sample ARL 4 at 2 m.
1.5-2 m.

C Softer, finer grained limestone with few obvious fossils, some large forams. Sample ARL 2 at 1.5 m.
1.3-1.5 m.

B Hard, finely crystalline limestone with neolaganids, *Oligopygus haldemani*, a cidaroid, ectoprocts, large forams, and other unidentified fragments. Sample ARL 1 at 0.7 m.
0-1.3 m.

INTERVAL OF UNKNOWN THICKNESS FLOODED.

A Boulders dredged from below water level. Coarse quartz grains in coarsely granular, hard limestone. *Periarchus lyelli*, large oysters, ribbed scallops, and other molluscs common. Fragments of other echinoids. Sample ARL X.

Appendix 2. Point count data (n=100-200 per slide depending on quality). Numbers represent percentages. Column headings are: A - carbonate mud, B - cement, C - peloids, D - echinoderms, E - ectoprocts, F - scallops, G - other molluscs, H - large Foraminifera, I - small Foraminifera, J - quartz, K - unidentifiable fossils, L - neomorphic spar, M - arthropods (?crabs), and N - red algae. Prefixes of samples represent: LQ - Leesburg Quarry (location b), OQ - Oakland Quarry (location c), ARL - Arlington Quarry (location e).

SAMPLE	A	B	C	D	E	F	G	H	I	J	K	L	M	N
LQ .4	32	8	13	5	32	2	-	2	4	-	2	-	-	-
LQ 1	2	28	12	10	26	5	4	1	5	-	7	-	-	-
LQ 1.8*	-	-	-	-	-	-	-	-	-	-	-	100	-	-
LQ 2.8*	14	-	2	13	8	10	-	3	4	-	-	46	-	-
LQ 3.6	24	17	7	22	15	5	1	3	1	-	5	-	-	-
LQ 4.5*	12	-	4	3	20	-	-	-	5	-	6	44	-	-
OQ X	-	25	8	-	2	3	1	-	1	58	2	-	-	-
OQ 1	4	26	30	6	5	2	3	15	5	-	4	-	-	-
OQ 2*	-	-	-	-	-	-	-	-	-	-	-	100	-	-
OQ 3	50	3	15	4	15	2	-	4	3	-	4	-	-	-
OQ 4	-	-	-	-	-	-	-	-	-	-	-	-	-	-
ARL X	-	20	33	2	3	5	14	11	-	8	4	-	-	-
ARL 1	-	23	22	4	9	-	2	28	5	-	7	-	-	-
ARL 2	6	21	17	9	13	-	2	9	2	-	2	15	2	2
ARL 3	-	16	20	13	8	-	-	15	8	-	4	16	-	-
ARL 4	-	20	28	3	6	1	4	21	5	-	6	-	-	6
ARL 5	10	22	36	1	1	1	1	21	1	-	4	-	-	2

* All samples are at least partially replaced with microspar which shows ghosts of scattered ectoprocts and forams. Echinoderm ossicles are also replaced, but retain their optical continuity so they appear to be coarser crystals. Assuming aggrading neomorphism, they probably had a muddy carbonate matrix originally.

NONGLACIAL LACUSTRINE VARVES FROM THE BENWOOD LIMESTONE (PENNSYLVANIAN) OF WEST VIRGINIA

DANIEL D. PETZOLD *Department of Geology, Indiana University Bloomington, IN 47405*

ABSTRACT

A thin bed of rhythmically laminated carbonate occurs in the lacustrine Benwood Limestone at Clarksburg, West Virginia. Laminae are arranged in organic/carbonate couplets that are interpreted to be nonglacial lacustrine varves. Application of the term "varve" is based largely on what is known of Benwood paleogeography and on comparison with examples of Holocene and ancient nonglacial varves. Average thickness of 72 of these varves is 1.0 mm (standard deviation 0.6 mm), indicating that the depositional rate during varve formation was on the order of 1 mm per year.

Seasonal phytoplankton blooms, probably associated with seasonal rainfall, are the likely mechanism of varve formation. In addition to seasonality, the presence of varves suggests that, at times during its history, the Benwood lake became relatively deep and stratified.

INTRODUCTION

A varve is a sedimentary layer deposited during one year. Varves usually consist of seasonally deposited laminae that are arranged in couplets (O'Sullivan, 1983). Lacustrine varves are important microstratigraphic features that can be used in the interpretation of paleoclimatology, paleoecology, and paleolimnology (Anderson and Dean, 1988). Unfortunately, examples of lacustrine varves are relatively rare in the pre-Cenozoic geologic record. Many examples of Late Paleozoic varves, for example, are marine in origin (Anderson and Dean, 1988) and consist of clastic/organic couplets or clastic/evaporite couplets (Bradley, 1931). Late Paleozoic lacustrine varves are restricted to clastic sequences associated with glacial environments (David, 1922; Hovey, 1924; Caldenius, 1938; Banks and others, 1954; Irving, 1957).

The purpose of this paper is to describe a rhythmically laminated carbonate sequence within the lacustrine Benwood Limestone (Pennsylvanian; Virgilian) of West Virginia. The laminae are interpreted to be lacustrine varves, and they are, to my knowledge, the first example of Late Paleozoic nonglacial lacustrine varves to be described. Although they constitute a minor part of the overall thickness of the Benwood Limestone, they provide further evidence for Pennsylvanian seasonality, provide an indication of depositional rate of Pennsylvanian organic-rich lacustrine carbonates, and provide evidence for the morphometry of the Benwood lake during its maximum development.

GEOLOGIC SETTING

The Dunkard Basin is a relatively small basin that trends northeast-southwest across Pennsylvania, Ohio, and West Virginia and contains marine and nonmarine sedimentary rocks that accumulated during Paleozoic time. During Early and Middle Pennsylvanian time, periodic marine transgressions entered the Dunkard Basin through a narrow seaway that connected the basin to more open marine conditions in the midcontinent. The Ames Limestone (Figure 1) was deposited during the last of these transgressions, after which deltaic sediments apparently prograded across the seaway, isolating the basin from further marine influence (Donaldson and Shumaker, 1979; Figure 2). Thus, Paleozoic rocks above the Ames Limestone consist of deltaic and fluvial siliciclastics and lacustrine carbonates. An abundance of nonmarine ostracodes, bivalves, gastropods, and fish combined with a complete lack of marine organisms in the lacustrine carbonates supports a freshwater origin for these rocks. Relatively high strontium isotope ratios in the limestones also indicate a freshwater depositional environment (Vasilkovas, 1987). The Benwood Limestone, the thickest of these lacustrine

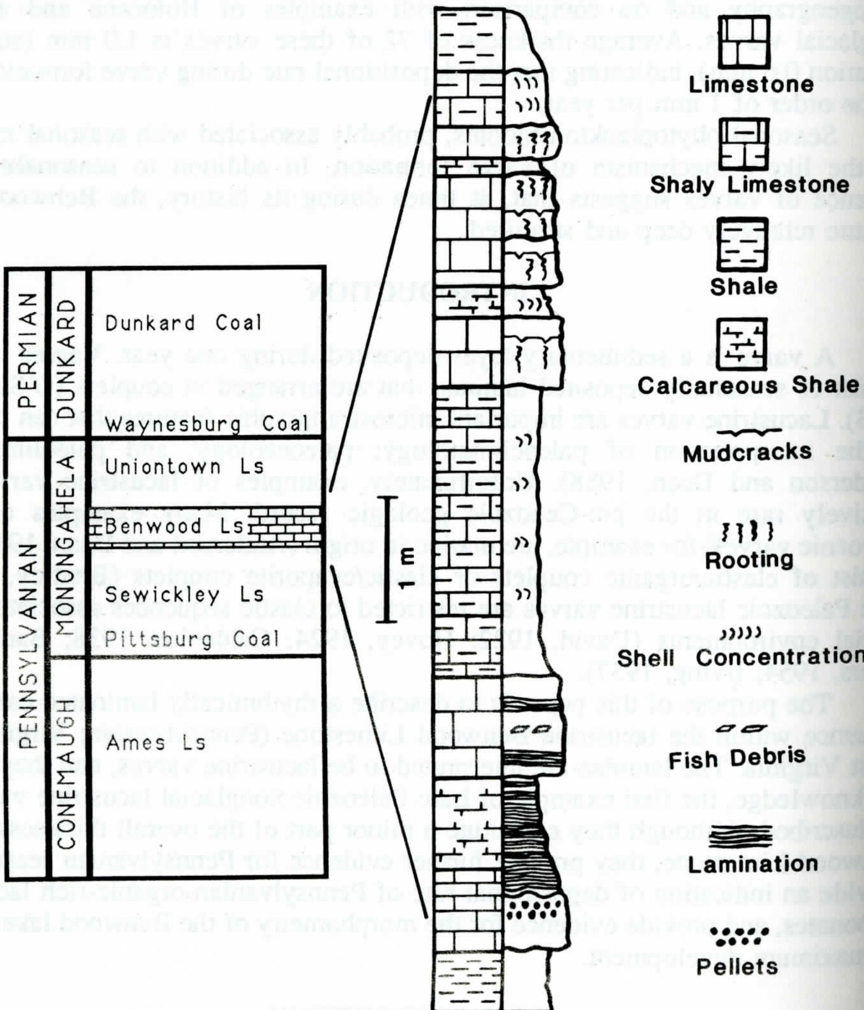


Figure 1. Stratigraphic setting of the Benwood Limestone in the Dunkard Basin and measured section of study interval.

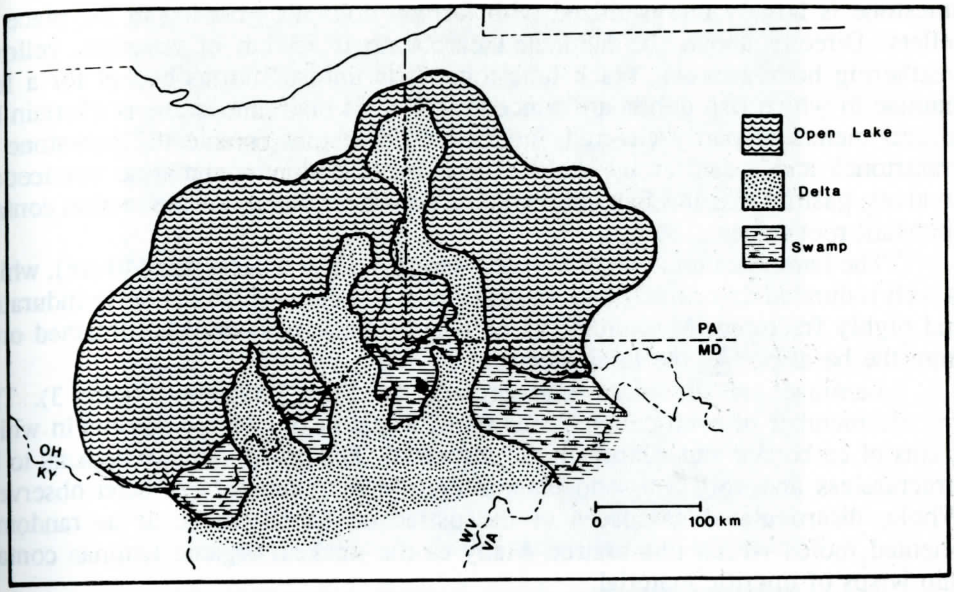


Figure 2. Map showing paleogeography of the Dunkard Basin at the time of deposition of the Monongahela Group (modified from Donaldson and Shumaker, 1979). Clarksburg is indicated by large circle.

carbonates is composed of thin beds of argillaceaous limestone. Carbonate beds range from mudstone to ostracodal or intraclastic grainstone and are interbedded with gray-to-green mudstone and shale (Berryhill, 1963).

The basin in which the Benwood Limestone was deposited probably consisted of numerous interdeltaic bays or small delta-plain lakes separated by elongate deltas that prograded towards the center of the basin (Donaldson and Shumaker, 1979; Figure 2). Subsurface data (Streib and Donaldson, 1969; Linger, 1979) indicate that the Benwood Limestone was not deposited as a basin-wide sheet but instead was deposited in pod-like bodies that presumably reflect interdeltaic, clastic-starved settings. Comparison of Benwood sections measured at widely spaced localities indicate that the lithologic sequence within the unit varies considerably. Local conditions within the interdeltaic bays or lakes probably influenced sedimentary patterns more than basin-wide conditions.

During Late Pennsylvanian time, the Dunkard Basin was situated at approximately 10 degrees south latitude (Scotese and others, 1979). The climate was warm temperate to tropical, but numerous episodes of lake desiccation indicate that the climate was not continuously humid (Berryhill, 1963; Cross, 1975).

DESCRIPTION OF THE LAMINATED ROCKS

A 1.1 m thick bed of dark, laminated carbonate occurs in the Benwood Limestone exposed in a roadcut of U.S. Route 50 at Clarksburg, West Virginia (mile 114.3 of Fedorko, McClelland, and Norton, 1979; Figure 2). At this locality, much of the Benwood is covered, and the laminated bed can be found only by excavation. Lateral persistence of the bed is indicated by its occurrence in a second outcrop on a side road approximately 1 km to the north east. Directly below the laminated bed is 35 cm of more resistant light-colored, peloidal limestone. This limestone is largely recrystallized with ferroan dolomite growing in the place of pellets. Directly above the laminated carbonate is 85 cm of resistant, yellow-weathering homogeneous, black limestone. It is unfossiliferous except for a few laminae in which fish debris are concentrated. This black limestone is overlain by several meters of partly covered, interbedded limestone (ostracodal packstone to grainstone) and fossiliferous shale. These rocks contain nonmarine ostracodes, bivalves, gastropods, and fish debris. Some beds near the top of the section contain abundant root traces.

The laminated unit can be divided into two parts: a basal part (40 cm), which is well indurated and brittle, and an upper part (70 cm), which is poorly indurated and highly fractured. Measurements of couplet thickness could be obtained only from the basal part of the laminated unit.

Laminae are organized into organic/carbonate couplets (Figure 3). The organic member of each couplet consists of a matrix of organic detritus in which grains of carbonate and quartz silt are dispersed. The organic matter appears to be structureless and very fine—no macroscopic plant remains have been observed. Whole, disarticulated carapaces of the ostracode *Carbonita* occur as randomly oriented molds within this matrix. Many of the thickest organic laminae contain thin wisps of micritic material.

The carbonate component of the laminated unit is concentrated in micritic laminae, most of which are extremely thin, faint, and discontinuous. X-ray diffraction analysis of whole rock samples indicates the carbonate is dominantly calcite with a small amount of ankerite. Illite, micas, or both occur in the matrix and appear to be concentrated in the carbonate laminae. Some carbonate laminae are unfossiliferous, but carapaces of *Carbonita* are abundant in most. They occur as whole, randomly oriented single valves. In many laminae, micrite is restricted to coatings on the ostracodes, giving the laminae a wavy, discontinuous appearance (Figure 3b). Small (~1 mm) pyrite framboids are common in some micritic laminae, but are rare to absent in the organic laminae. Contacts between organic laminae and micritic laminae are generally sharp and straight except where ostracodes are abundant.

Thickness of 72 consecutive couplets was measured using a binocular microscope and micrometer. The average thickness of measured couplets is 1.0 mm (thickest, 3.5 mm; thinnest, 0.3 mm) with a standard deviation of 0.6 mm. Autocorrelation of the measured couplets shows no statistically significant cyclicity (Figure 4). The relationship between neighboring laminae varies through the section. Throughout most of the bed, organic laminae comprise 70 to 80 percent of the thickness of each couplet, although thin intervals occur in which micritic laminae are of equal or greater thickness than their bounding organic laminae.

Several thick (4-5 mm) laminae are present within the laminated sequence (Figure 3a). They are composed of either a homogeneous mixture of organic material and carbonate similar to the material in the surrounding laminae or concentrations of ostracode carapaces, fish debris, and quartzose silt. The base of the thicker laminae is erosional, cutting across several couplets and containing intraclasts. One such erosional horizon has been tentatively correlated between the two outcrops in which the laminae occur. The relative position of this bed in the two outcrops indicates that at least 3 cm of section is missing from the northeast section.

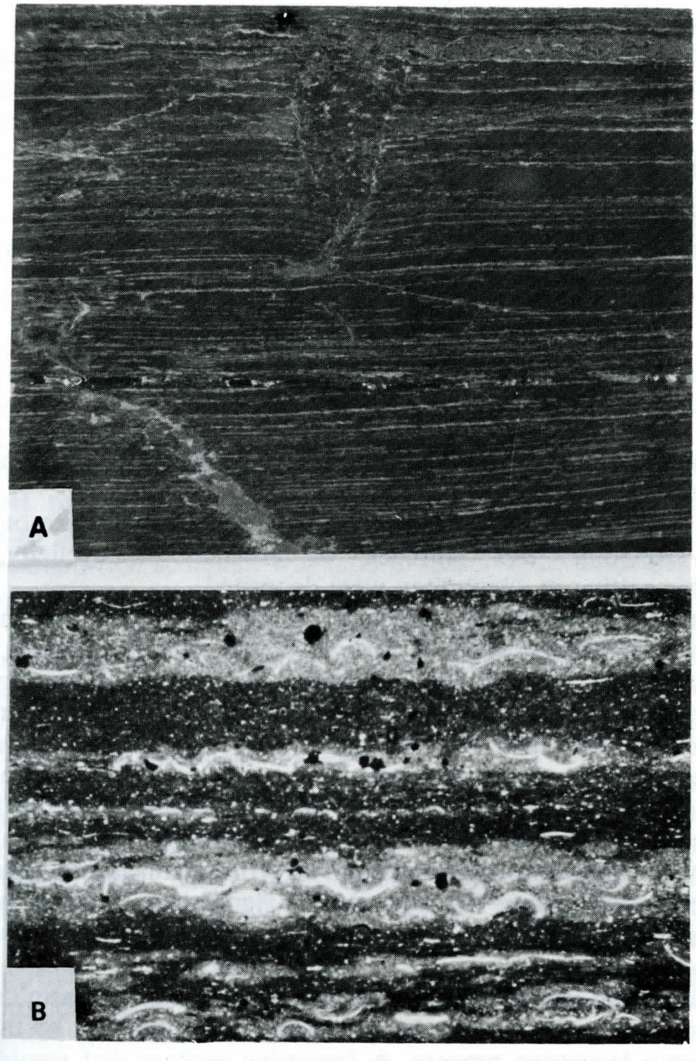


Figure 3. a. Portion of slabbed section of laminated limestone. Note discontinuous nature of micritic laminae and coarse-grained (turbidite?) lamina at top. Field of view is 7.5 cm in width. b. Photomicrograph (35x) of laminated limestone. Note concentration of ostracodes and pyrite in micritic laminae.

INTERPRETATION

Rocks composed of alternately repeating laminae of two different kinds of material are called rhythmites. They result from either regular changes in transport of material or regular changes in production of material (Reineck and Singh, 1980). The former mechanism occurs in the form of short-term variations in flow characteristics (i.e. tidal currents or turbidity currents) whereas the latter is more often in the form of long-term changes such as seasonality (i.e. regular variations in temperature or rainfall).

The rhythmic laminae described above, then, were produced as a result of either short-term or long-term changes in the depositional environment.

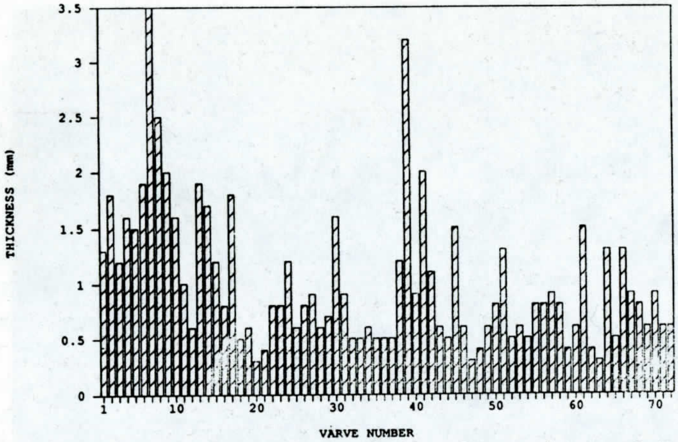


Figure 4. Graph of thickness for 72 carbonate and organic couplets near base of laminaed unit.

Interpretation must rely chiefly on their composition, the presence or absence of sedimentary structures, the presence or absence of cyclicity, and the paleogeographic setting in which they were deposited.

Rhythmites produced by short-term changes in flow characteristics are almost always characterized as being clastic in nature and as containing a variety of sedimentary structures such as climbing ripples and normal or herring-bone cross lamination. In addition, some of these rhythmites (i.e. tidalites) exhibit marked cyclicity. Laminae produced by short-term flow fluctuations are arranged in couplets of coarse-grained and fine-grained members that have been segregated due to differences in settling rates of these fractions under different flow conditions. As a result, these laminae are often graded. Both members of the couplets from the Benwood Limestone are very fine grained, are nonclastic, lack sedimentary structures, show no statistically significant cyclicity, and are not graded. This suggests strongly that they did not originate as a result of a short-term fluctuation in flow parameters.

In addition, the paleogeographic setting of the Dunkard Basin during Upper Pennsylvanian time suggests that the mechanism of formation of the Benwood rhythmites was a lacustrine process. A lacustrine process that would alternately produce organic-carbon-rich sediment and carbonate-rich sediment on a millimeter scale is seasonal sedimentation. Recognition of ancient varves is based largely upon analogy with structures, thicknesses, and textures of modern varved sediments (Bradley, 1937; Anderson and Dean, 1988). Apparently, the most reliable characteristic of nonglacial varves is the presence of organic matter in one lamina of a couplet (Bradley, 1937). In terms of their composition, thickness, and general appearance, the organic/carbonate couplets described in this paper compare well with organic/carbonate varves described from Holocene (Dean and Fouch, 1983; O'Sullivan, 1983) and from ancient (Dean and Fouch, 1983; Anderson and Dean, 1988) lacustrine sediments in terms of composition, thickness, and general appearance.

The term "varve" clearly denotes annual deposition, and one may question whether the term can be applied with certainty to ancient rocks. Intra-annual laminations can be produced in lacustrine sediments by high winds or heavy rains (Saarnisto, 1986), and the presence of longer-than-annual phytoplankton blooms in Holocene tropical African lakes (Melack, 1979) suggests that it may be possible to produce inter-annual laminations. Normally, the annual nature of rhythmites must be established before the term "varve" can be applied to Holocene sediments. For Paleozoic-age rocks, however, this is much more difficult. The comparison of thickness, structure, and components between the Benwood rhythmites and Holocene varved sediment, however, probably justifies application of the term "varve" to the Pennsylvanian laminations. Some couplets may represent shorter- or longer-than-annual periods, but the persistence of these millimeter-scale rhythmites is best explained by regular seasonality.

DISCUSSION

Formation of organic/carbonate varves in nonglacial lacustrine environments may occur when seasonal phytoplankton blooms act to temporarily raise the pH of lake water through uptake of CO_2 (O'Sullivan, 1983). This induces precipitation of calcite if the water is near saturation with respect to the mineral. After the pH returns to its previous level, a lamina rich in organic matter accumulates as phytoplankton settle out of the water column. Acids produced by the organic-carbon-rich sediment will tend to dissolve the subsequently deposited carbonate lamina, and micrite will be preserved only when its rate of deposition exceeds the system's ability to dissolve it (Dean and Fouch, 1983). Years in which carbonate production was low may not be preserved in the rock record, thus thicker organic laminae may record two or more years of sedimentation. Alternatively, varves may be produced where seasonal temperature fluctuations periodically induce precipitation of carbonate minerals.

Given the tropical setting of the Dunkard Basin during Paleozoic time, seasonal temperature fluctuations are not likely to have been great. However, seasonal phytoplankton blooms in Holocene tropical and warm-temperate African lakes do result from either hydrological features (water input-output) or hydrographic features (water-column circulation; Talling, 1986). Hydrographic seasonality is unlikely to produce varves because water-column circulation will tend to resuspend bottom sediments, and circulation will act to oxygenate bottom water and promote decay of organic matter and bioturbation of bottom sediment. Seasonal rainfall (hydrologic seasonality), however, would not necessarily create bottom currents or disrupt lake stratification and is thus more likely to create preservable varves in a tropical setting. The varves in the Benwood Limestone, then, are probably the result of seasonal rainfall, which promoted phytoplankton blooms in the Benwood Lake. This conclusion is supported by other evidence for seasonal rainfall in equatorial regions during the late Paleozoic (Zangerl and Richardson, 1963; Cross, 1975; Broadhurst and others, 1980; Cecil and others, 1985).

Seasonality may also be reflected in the fact that *Carbonita* is largely restricted to the calcitic portion of the couplets. Most Holocene North American freshwater ostracodes display a seasonality that is probably related to temperature fluctuation (Pennak, 1953). Little evidence exists for seasonal temperature fluctuations in equatorial regions during the Pennsylvanian. Instead, seasonal growth of *Carbonita* may have been a response to seasonally available resources or to seasonal variations substrate conditions.

Although lakes where varves have been described are on average less than 1 km in length, have a surface area of less than 20 ha, and are over 15 m deep, many exceptions are known, and varves are also found in isolated deep basins within large lakes with hundreds of square kilometers of area (Saarnisto, 1986). It was probably the latter setting—a localized deep area within the Benwood lake—in which the Benwood varves were deposited. The bottom of this body of water was flat enough to prevent slump and frequent turbidity currents, and it was below storm wave base (O'Sullivan, 1983). The lake was probably thermally stratified and lacked seasonal mixing. Bottom water was sufficiently depleted in oxygen to prevent oxidation and bioturbation of the organic-rich sediment (Saarnisto, 1986).

The overall sedimentologic history of the study interval at Clarksburg records an apparent shallowing-upwards lacustrine sequence. The pelleted limestone, varves, and black limestone indicate a relatively deep, sublittoral depositional environment. The transition from the pelleted limestone to the varve sequence occurred when the lake became stratified and the bottom waters became anoxic. The thicker, erosional laminae in the varve sequence are probably a result of turbidity currents that scoured previously deposited sediment or introduced allochthonous sediment. These events may have removed the sedimentary record of many years, however it is impossible to know how much sediment has been removed. Variations in varve thickness resulted from year-to-year variability of water chemistry, production of autochthonous material, and climate. The upwards increase in fissility of the varved unit probably resulted from a slowly increasing rate of silt influx and dominance of organic production over carbonate production. The main difference between the environment that produced the varved unit and that which produced the overlying black limestone was the presence of bottom currents or soft-bodied organisms that homogenized the sediment. This probably reflects shallowing, loss of the lake's stratification, or both. The sparseness of fossils in the black limestone, however, suggests that this unit was deposited in sublittoral water that may have been anoxic. Above the black limestone, the sequence of interbedded packstones, grainstones, and fossiliferous shales indicates deposition in shallow, oxygenated water that was affected by wave or current action and influxes of terrigenous silt. Rooting near the top of the section suggests periodic desiccation of the lake and exposure of the lacustrine carbonate mud.

The average thickness of 1.0 mm for couplets in the interval studied suggests that if gaps are absent, the rate of deposition was on the order of 1 mm/year. Therefore, conditions necessary for deposition and preservation of these varves existed for at least 1,100 years.

CONCLUSIONS

1. Rhythmic laminae that occur in a bed of the Benwood Limestone at Clarksburg, West Virginia are interpreted as nonglacial lacustrine varves on the basis of the paleogeographic setting of the Dunkard Basin and by comparison with modern and ancient examples of lacustrine varves. Millimeter-scale alternation of calcitic micritic and organic-C-rich laminae is the best evidence that these are nonglacial varves. A variety of negative evidence, such as the lack of clastic material, size sorting, cyclicity, and current-generated sedimentary structures, argues against alternate interpretations such as tidalites and turbidites.
2. The presence of varves indicates that the lake in which the Benwood Limestone was deposited was essentially flat-bottomed, deep, stratified, and anoxic during at least part of its history.
3. Couplets of organic matter and calcite indicate seasonal phytoplankton blooms. These blooms were probably related to seasonal rainfall and water influx as

opposed to seasonal lake circulation, which would have stirred the sediment and oxygenated the bottom water. The preferential occurrence of *Carbonita* in micritic laminae suggests that the ostracode had a seasonal life cycle.
4. The average thickness of 72 measured couplets is 1.0 mm, indicating that the depositional rate for the combination of organic matter and inorganic calcite was approximately 1 mm/year.

ACKNOWLEDGEMENTS

This paper benefitted greatly from critical comments made by Mark A. Brown, Roger Y. Anderson, and Doug Kirkland. Research was supported in part by a grant from the Department of Geology at Indiana University.

REFERENCES CITED

Anderson, R. Y. and Dean, W. E., 1988, Lacustrine varve formation through time: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 62, p. 215-235.

Banks, M. R., Lovelady, J., and Scott, D. L., 1954, First record of Permian varves in Tasmania: Australian Journal of Science, v. 17, p. 66-67.

Berryhill, H. L. Jr., 1963, Geology and coal resources of Belmont County, Ohio: U. S. Geological Survey Professional Paper 380, 113 p.

Bradley, W. H., 1931, Non-glacial marine varves: American Journal of Science, 5th ser., v. 22, p. 318-330.

Bradley, W. H., 1937, Non-glacial varves, with selected bibliography: National research Council Report of the Committee on the Measurement of Geologic Time, p. 32-43.

Broadhurst, F. M., Simpson, I. M., and Hardy, P. G., 1980, Seasonal sedimentation in the upper Carboniferous of England: Journal of Geology, v. 88, p. 639-651.

Caldenius, C. C., 1938, Carboniferous varves measured at Paterson, New South Wales: Geologiska Foreningens Stockholm Forhandlingar, v. 60, p. 349-364.

Cecil, C. B., Stanton, R. W., Neuzil, S. G., Dulong, F. T., Ruppert, L. F., Pierce, B. S., 1985, Paleoclimatc controls on Late Paleozoic sedimentation and peat formation in the central Appalachian Basin (U.S.A.): International Journal of Coal Geology, v. 5. p. 195-230.

Cross, A. T., 1975, Discussion of paper by Aureal T. Cross, *in* Barlow, J. A., ed., The age of the Dunkard, Proceedings of the First I. C. White Memorial Symposium: West Virginia Geological and Economic Survey, p. 300-304.

David, T. W. E., 1922, The "varved shales" of Australia: American Journal of Science, v. 3, p. 115-116.

Dean, W. E. and Fouch, T. D., 1983, Lacustrine environment, *in* Scholle, P. A., Bebout, D. G., and Moore, C. H., eds., Carbonate depositional environments: American Association of Petroleum Geologists Memoir 33, p. 98-130.

Donaldson, A. C. and Shumaker, R. C., 1979, Late Paleozoic molasse of central Appalachians, *in* Donaldson, A. C., Presley, M. W., and Renton, J. J., eds., Carboniferous Coal Short Course and Guidebook: Volume 3: West Virginia Geological and Economic Survey Bulletin B-37-3, p. 1-42.

Fedorko, N., McClelland, S. W., and Norton, C. W., 1979, Road log—sixth day,

in Englund, K. J., Arndt, H. H., Henry, T. W., eds., Proposed Pennsylvanian System stratotype Virginia and West Virginia, Field trip no. 1: Ninth International Congress of Carboniferous Stratigraphy and Geology, p. 55-64.
Hovey, E. O., 1924, Discussion (Note on varved Permo-Carboniferous slates in Australia): Geological Society of America Bulletin, v. 35, p. 67.
Irving, E., 1957, Direction of magnetization in the Carboniferous glacial varves of Australia: Nature, v. 180, p. 280-281.
Linger, D. B., 1979, Stratigraphic analysis of the lower Monongahela Groups coals and associated rocks (Pittsburgh-Redstone Interval) in Marion and Monongalia Counties, West Virginia and Greene County, Pennsylvania, *in* Donaldson, A. C., Presley, M. W., and Renton, J. J., eds., Field trip guidebook for Carboniferous coal short course: West Virginia Geological and Economic Survey Bulletin B-37-2, p.93-120.
Melack, J. M., 1979, Temporal variability of phytoplankton in tropical lakes: Oecologia, v. 44, p. 1-7.
O'Sullivan, P. E., 1983, Annually laminated lake sediments and the study of Quaternary environmental changes—A review: Quaternary Science Reviews, v. 1, p. 245-313.
Pennak, R. W., 1953, Fresh-water invertebrates of the United States: Ronald Press, New York, 769 p.
Reineck, H.-E. and Singh, I. B., 1980, Depositional sedimentary environments: Springer-Verlag, New York, 549 p.
Saarnisto, M., 1986, Annually laminated lake sediments, *in* Berglund, B. E., ed., Handbook of Holocene palaeoecology and palaeohydrology: John Wiley & Sons, New York, p. 343-370.
Scotese, C. R., Bambach, R. K., Barlon, C., and Ziegler, A. M., 1979, Paleozoic base maps: Journal of Geology, v. 87, p. 217-277.
Streib, D. L. and Donaldson, A. C., 1969, Stratigraphy of the Monongahela Group in the Wheeling, West Virginia-Ohio 15-minute quadrangle, *in* Donaldson, A. C., ed., Some Appalachian coals and carbonates: models of ancient shallow-water deposition: Geological Society of America Field Trip Guidebook p. 123-156.
Talling, J. F., 1986, The seasonality of phytoplankton in African lakes: Hydrobiologia, v. 138, p. 139-160.
Vasilkovs, I. N., 1987, 87Sr/86Sr and geochemical variations in the Pennsylvanian and Permian nonmarine and marine limestones of southeastern Ohio, northwestern West Virginia, and southwestern Pennsylvania: Master's thesis, Miami University, Oxford, Ohio, 96 p.
Zangerl, R. and Richardson, E. S., 1963, The paleoecological history of two Pennsylvanian black shales: Fieldiana Geology Memoir 4, 352 p.

A COMPARISON OF SUSPENSATE AND BOTTOM STREAM SEDIMENT GEOCHEMISTRY AT A PB OCCURRENCE IN THE SHENANDOAH VALLEY ZN DISTRICT, NORTHWEST VIRGINIA

FREDERIC R. SIEGEL

Department of Geology
The George Washington University
Washington, D.C. 20052 USA

ABSTRACT

Suspended sediment was collected from stream water together with equivalent active sediment from the stream bottom at 42 sites near the Gordon Pb property, Shenandoah County, Virginia (USA) to test the relative effectiveness of suspended and bottom sediment in the geochemical exploration for stratabound Mississippi Valley type ores. Zn was a pathfinder to the mineralization, and Zn in the suspensates gave a larger number of strong anomalies, higher in magnitude, than Zn in the −100 mesh size of the bottom sediment, either as total Zn or as Zn extracted with cold 3% HCl. Pb was detected only in the cold acid extraction. In combination with cold acid extracted Zn and Cu, Pb formed a multielement anomaly that located the Gordon property mineralization and had a longer downstream dispersion than did the suspensate or total dissolution multielement anomalies. Groundwater inflow or spring discharge from the mineralized host rock into the stream may influence the development of and locations of the anomalies. Both suspensates and cold acid extracted samples are more effective for geochemical prospecting in this area than the total dissolution samples of the −100 mesh size of the active stream sediments.

INTRODUCTION

Suspended sediment is transported by flowing water and consists of both mineral (inorganic) and organic phases. The suspended mineral matter is from physical weathering and erosion of particulates into the drainage, and from precipitation reactions in the hydrogeochemical environment. The geochemistry of suspended sediments reflects the geochemistry of clastic material eroded from a drainage basin and thus can be used in the same way as stream bottom sediments in exploration geochemistry. In addition, the suspended sediment by sorption of trace metal, may indicate the chemistry of the immediate aqueous environment fed in part by inflow or spring discharge of groundwater. Thus, the suspensate may be a truer representative sample for geochemical prospecting than a specific size fraction of the bottom sediment in which redox changes and other diagenetic effects can alter the geochemistry. This study was done to test the relative effectiveness of the geochemistry of suspended and bottom sediments on the exploration for stratabound Mississippi Valley type ores.

The suspended sediment has had limited testing as a geochemical prospecting sample mainly because of the difficulty in obtaining a sample (without using on-site, heavy, electrical vacuum-filtering systems), but also because of successful use of the stream bottom sediment.

Previous Work

Perhac and Whelan (1972) made a three sample study on suspended solids in drainage from known mineralization in the northeast Tennessee zinc district with one sample 1200 m upstream from the mineralization and samples at 35 m and 1000 m downstream from it. The contents of Zn and Pb in the colloidal fraction (0.01 - 0.15 µm) "found" the mineralization as did the Zn and Pb in a coarser (>0.15 µm) suspended fraction. However, the authors reported that so much water (40 liters per sample) had to be transported to the laboratory for time-consuming ultra-centrifuging to separate the suspended sediment from the water, that the use of suspended sediment for prospecting was impractical. Perhac and Whelan (1972) also found that the >5 µm size fraction of the bottom sediment contained high amounts of Zn downstream from the mineralization but the magnitude of the highs was considerably less than that shown by the colloidal fraction.

Carserud (1983) studied stream suspended sediment from the watershed of the Verkaan River (southeast Sweden) which drains a black shale sequence. He found that elevated contents of As, Sb, V and Pb in the suspended matter showed enrichment of these metals over their contents in the black shale by 25× for As, 10× for Sn, 7× for V, and 2× for Pb. This increased metal concentration is precisely the relationship that is the basis for using stream suspended sediment in exploration geochemistry.

As already noted, one major drawback to the evaluation of and then possible use of suspended stream sediment for geochemical exploration was the problem of collecting the sample in situ especially in rather inaccessible field areas and in areas with difficult terrain where vehicle- or boat-mounted electrical vacuum-filtering systems can not be used. These are the areas where increased geochemical exploration will take place in the future. To overcome this sampling problem, a light-weight, polypropylene, mechanical (hand-operated) vacuum-filtering unit was developed and used in a study on suspended sediments collected at six sites in drainage from a gold occurrence in Maryland (Siegel, 1982). Comparisons of the geochemistries of the suspended matter (>0.45 µm) and the −100 mesh size fraction of the corresponding bottom sediment showed that Zn and Cu in the suspended mineral matter defined the position of the gold mineralization but that the geochemistry of the fine size fraction of the bottom sediment did not.

There are many factors that may affect the geochemistry of suspended sediments for use in geochemical prospecting or other research programs. For example, what is the seasonal effect, if any, on suspended sediment metal content? Burman (1983) found that during a 24 month sampling at six sites in the Kalix River (northern Sweden), Mn and Ba had concentration maxima late in the summer (August). He believed that this might be related to a biogeochemical tendency for Mn and Ba to be selectively enriched in forests during the winter, and their subsequent introduction into the hydrogeological regime. Fe had an invariant correlation (−0.97) with Mn. The data for Zn, Cu, Ni and Cr during the same period were not reported in the same detail as for Mn, Ba and Fe, but correlations between them and either Mn or Fe were not significant. This suggests a lack of temporal geochemical concentration relations. Other factors to be investigated include, for example, the response of mineral suspensates in different environments with similar mineralization, the applicability of mineral suspensates to locate different kinds of mineralization with their attendant different geologies, and the seasonal effect on element concentrations of suspensates in different climatic regimes.

MINERALIZATION AND LOCAL CONDITIONS

The Gordon property is located in the Timberville 7.5 quadrangle, Shenandoah County, Virginia (USA), about 7.4 km N13°E of the town of Timberville (Figure 1). The topography is gently rolling with a relief of about 60 m in the study area, which has a grass-tree vegetation, and water flowing from a slow to rapid rate in streams generally 0.3 to 0.9 m deep and 0.6 to 3 m wide. The climate is temperate and there was good water flow during the June, 1983, field period. Herbert and Young (1956) have described this and other properties in the Shenandoah Valley Zn district in some detail. They wrote that at the Gordon property, mineralization with galena and small amounts of sphalerite and pyrite occurs in the dolomite breccia in the upper Beekmantown Group. The Beekmantown strata strike N34°E and dip 30°SE. The galena occurs as irregular masses in vein dolomite and as thin fracture fillings in breccia blocks whereas the sphalerite occurs in the dolomite cement and replacing the outer surfaces of dolomite blocks. Eight to 24 m east of the mineralization, drilling intersected the breccia zone containing minor amounts of lead and zinc sulfides within about 11 m of the surface. A few hundred meters north of the Gordon property sparse outcrops contain disseminated sulfides in partially coarsely crystalline dolomite that may be related to the main breccia mineralization. Drilling about 24 m east of these northern outcrops revealed only minor traces of sulfides and these were within about 11 m of the surface. Results from this exploration did not warrant major development and the Gordon prospect was not mined. Because of the lack of development in this lightly populated agricultural area, contamination is not considered to be a problem for geochemical prospecting.

METHODOLOGY

Field Work

Suspended sediments and their bottom sediment correlatives were collected from 42 sites at 150 to 200 m intervals from the drainage associated with the Gordon property mineralization (Figure 1). This dense sample spacing was used in order to have sufficient data to find the maximum spacing that can effectively target the type of mineralization studied here. The samples were collected in numerical sequence but were ordered according to a random number scheme during the laboratory preparation, and reordered randomly again for the geochemical analyses. The entrained suspended matter was separated from the stream water in situ using a polypropylene suction nozzle funnel (Nalgene) with a polypropylene mechanical (hand-operated) vacuum pump (Siegel, 1985). About six liters of water were pumped through 142 mm diameter Millipore filters with 0.45 µm nominal pore size openings for each loading. Filters were loaded twice (to economize on the cost of filters) by filtering until funnel pores became clogged, releasing the vacuum, shifting the filter to a new position, and resuming pumping. The clogging means that the samples also contained a small amount of suspended matter smaller than 0.45 µm. Filters were transferred to samples bags with plastic forceps. Millipore filters were used instead of other commercially available ones because they are faster filtering and time is important for geochemical exploration programs.

An integrated sample of stream bottom sediment was collected after the suspended sampling was completed and represents about a 10 m run of the stream. Readings of streamwater pH were made with a portable pH meter. Thirty to 40 minutes were spent at each sampling site.

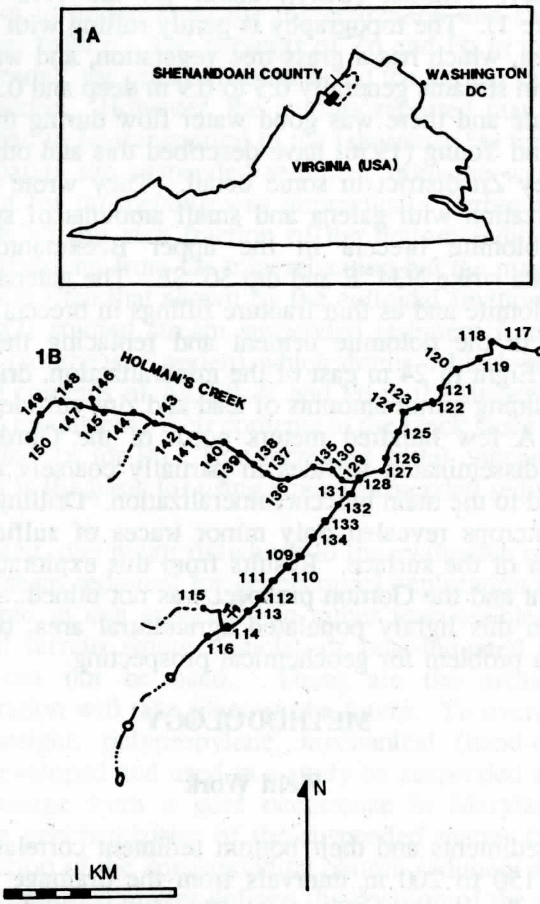


Figure 1. A. General location map. Black rectangle marks location of the Beekmantown Zn belt. B. Numbered sample locations in drainage associated with the Gordon property.

Laboratory Work

In addition to the mineral suspensate, a total dissolution of the –100 mesh size fraction of the stream bottom sediment, and a cold acid extraction of this fraction were prepared for geochemical comparison. These will be referred to later in the text as M-MS (metal-mineral suspensate), M-TD (metal-total dissolution), and M-CXT (metal-cold extractable).

The mineral suspensate was prepared by combusting the suspended sediment in a porcelain dish in a muffle furnace at 50° C increments up to 500° C (to remove organic material) and the combustion continued for 16 hours. The incremental heating eliminated the filter popping at 150° C-200° C reported by Burman (1983). Twenty-five mg of ashed mineral suspensate was mixed with 125 mg of lithium metaborate and fused in a pre-fired, tared, graphite crucible at 1000° C for 12

Table 1. Summary of geochemical data from mineral suspensates and associated bottom sediments (–100 mesh size fraction: total dissolution and cold 3% HCl extraction) from the Gordon property drainage system. Forty-two sites were sampled. Values are in ppm ounless otherwise noted.

	Zn	Cu	Fe%	Mn	Pb
Mineral Suspensates					
X	443.2	111.5	3.66	654.1	ND
s	197.8	35.7	0.68	277.1	
X + 1.1s	660.8	149.8	4.41	958.9	
X + 2s	838.8	182.9	5.02	1208.3	
Low value	152.0	50.0	1.75	87.0	
High value	1143.0	241.0	5.69	2136.0	
Total Dissolution					
X	127.4	99.0	2.55	588.4	ND
s	94.9	50.4	0.66	216.6	
X + 1.1s	231.8	153.4	3.27	826.7	
X + 2s	317.2	199.8	3.87	1021.6	
Low value	56.0	BDL	1.66	316.0	
High value	564.0	233.0	3.96	1120.0	

	Zn	Cu	Fe	Mn	Pb
Cold 3% HCl Extraction					
X	1.64	0.59	118.4	33.2	1.83
s	0.86	0.25	55.7	14.9	1.46
X + 1.1s	2.58	0.86	179.7	49.6	3.43
X + 2s	3.36	1.09	229.8	63.1	4.75
Low value	0.63	0.30	45.5	17.8	0.56
High value	4.50	1.20	281.2	73.3	8.44

ND = not detected
BDL = below detection level

minutes following the technique of Medlin and others (1969). The molten bead produced was transferred to a polypropylene beaker containing 25 ml of 3% HNO_3 for dissolution with magnetic stirring. Solution takes about 5 minutes. The sample was transferred to pre-cleaned plastic vials through filter paper (to remove graphite particles) for subsequent analysis.

Samples of the –100 mesh size fraction of the stream bottom sediment were totally dissolved following the technique of Medlin and others (1969). In addition, a cold acid extract was prepared by reacting 500 mg of the –100 mesh size fraction of the stream bottom sediment with 20 ml of 3% HCl on a shaking table for 20 minutes. The cold acid extract was filtered through Whatman 41 paper into plastic vials for analysis.

The samples were analyzed for Zn, Cu, Pb, Fe and Mn using flame atomic absorption. Blank-corrected standards were matrix matched to the sample type.

Mineralogy of the untreated suspended matter was determined using X-ray diffraction analysis.

Data Reduction

The analytical data were processed to determine mean and standard deviation values, frequency distributions, and cumulative frequency plots. Strong and moderate anomaly concentrations were arbitrarily established at values greater than X + 2s (mean plus two standard deviations) for the strong anomaly, and at values between X + 1.1s and X + 2s for the moderate anomaly. Correlation coefficients were calculated to give indication of interelement relationships.

RESULTS AND DISCUSSION

Element concentrations

The analytical results and selected statistical parameters are summarized in Table 1. Concentrations of all elements are higher in the mineral suspensates than the total contents of the –100 mesh size samples. This enrichment is especially pronounced for Zn and to a lesser degree for Fe, Cu and Mn. Only the cold 3% HCl extraction samples contain detectable Pb (greater than 0.5 ppm) but this could be expected since Pb will be mobilized during the leaching but can have its concentration diluted below the analytical detection limit by the other components in the mineral suspensates and the total dissolution suite.

Frequency Distributions

Frequency distributions for all elements in all sample types except for Fe-MS and Mn-MS have the positive asymmetry that is most common in geochemical populations. Fe-MS and Mn-MS have distributions that approximate normality. Bimodal or polymodal distributions (e.g., Zn-MS, Zn-TD, and Pb-CXT) with high value frequency classes suggest the influence of mineralization on the population. Figure 2 shows the frequency distribution for Zn-MS and the corresponding cumulative frequency graph which also indicates the presence of a high value population, possibly from the mineralization.

Geochemical Anomaly Maps and Downstream Metal Concentration Profiles

Geochemical maps with the locations of strong and moderate anomalies show that single and multielement anomalies in the three sample types studied clearly locate the Gordon property mineralization (Figures 3A, 4A, and 5A). If only the ore elements Pb and Zn are considered, plus Cu as a non-ore chalcophile associate, it is obvious that because of the Pb the CXT samples present a target with a larger number of strong and moderate anomalies and with a longer dispersion train (because of Cu) than the MS suite. The Pb, which was detected only in the CXT analyses, has about half the dispersion of Zn (strong and moderate anomalies) in either the MS or CXT samples. The TD group Zn and Cu are the least effective in the three sample types for highlighting the mineralization.

With Zn alone as the pathfinder element (Figure 6A), Zn-MS presents the best anomaly string in terms of number of strong anomalies, and has a dispersion

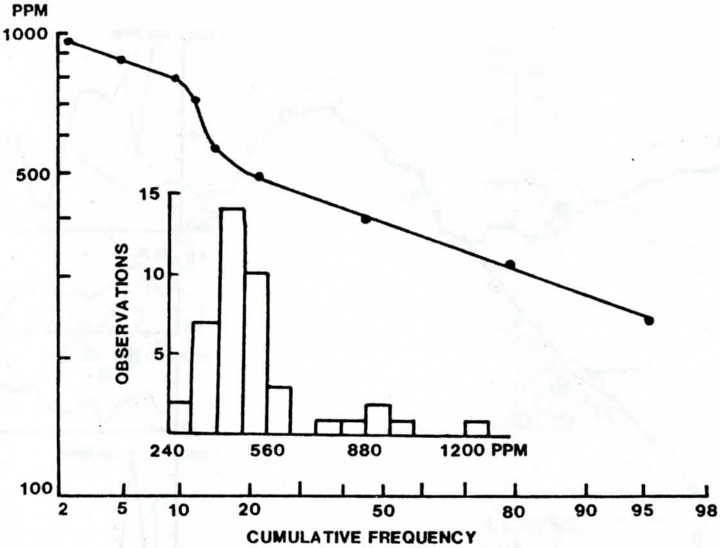


Figure 2. Frequency distribution and corresponding cumulative frequency plot for Zn in the mineral suspensates.

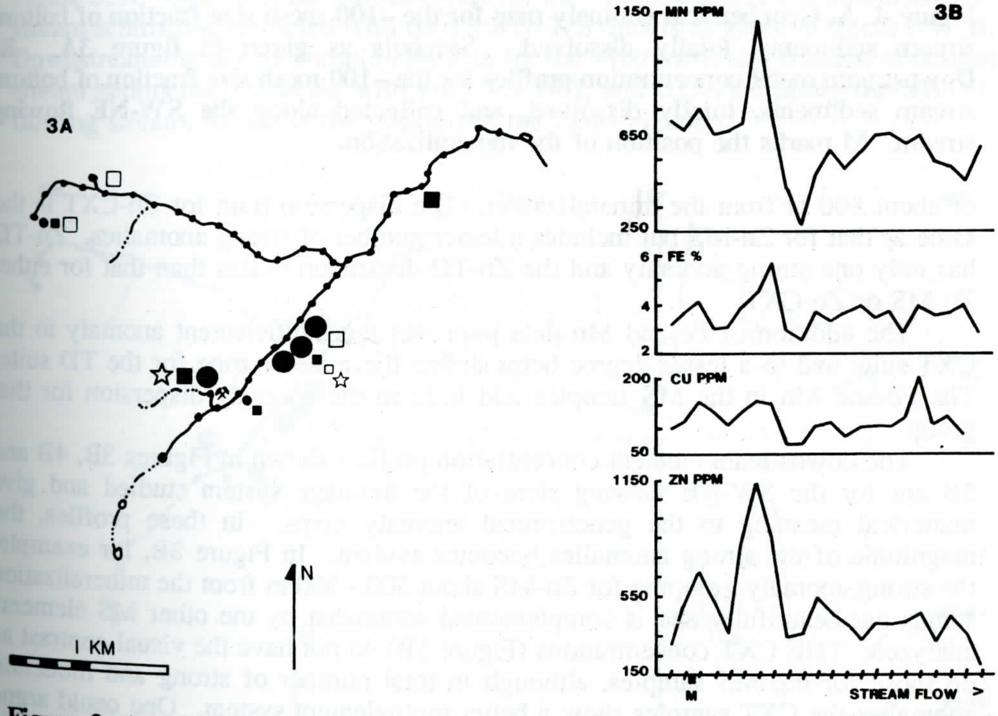


Figure 3. A. Geochemical anomaly map for suspensate samples. In this figure and figures 4A and 5A, large symbols represent strong anomalies, small symbols represent moderate anomalies; full circles = Zn; full squares = Cu; open squares = Fe; open stars = Mn; open circles = Pb. B. Downstream metal concentration profiles for suspensate samples collected along the SW-NE flowing stream. M marks the position of the mineralization.

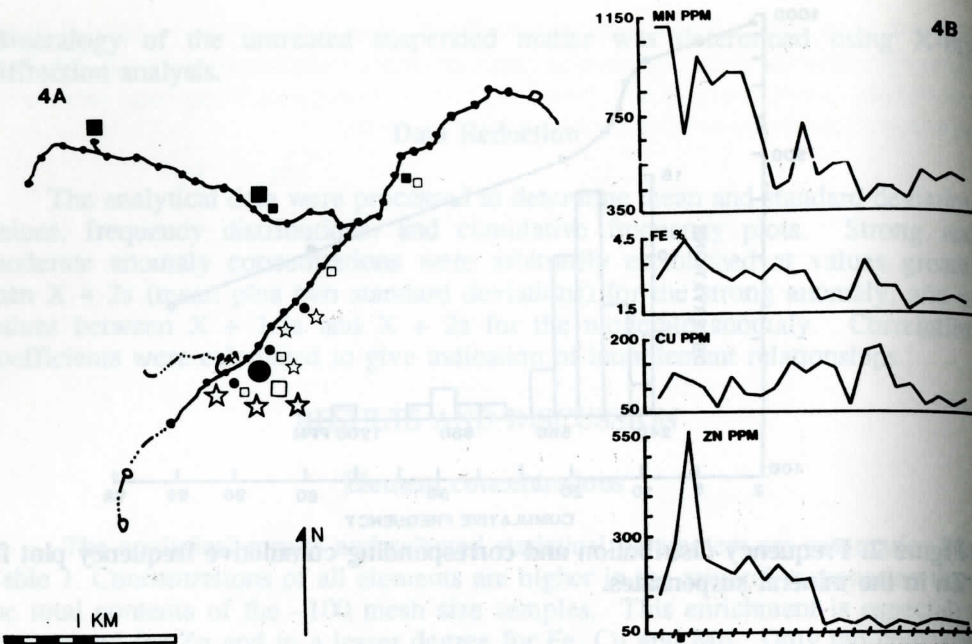


Figure 4. A. Geochemical anomaly map for the –100 mesh size fraction of bottom stream sediments, totally dissolved. Symbols as given in figure 3A. B. Downstream metal concentration profiles for the –100 mesh size fraction of bottom stream sediments, totally dissolved, and collected along the SW-NE flowing stream. M marks the position of the mineralization.

of about 800 m from the mineralization. The dispersion train for Zn-CXT is the same as that for Zn-MS but includes a lesser number of strong anomalies. Zn-TD has only one strong anomaly and the Zn-TD dispersion is less than that for either Zn-MS or Zn-CXT.

The addition of Fe and Mn data improves the multielement anomaly in the CXT suite and to a lesser degree helps define the mineral zone for the TD suite. The Fe and Mn in the MS samples add little to the anomaly dispersion for that group.

The downstream element concentration profiles shown in Figures 3B, 4B and 5B are for the SW-NE flowing stem of the drainage system studied and give numerical meaning to the geochemical anomaly maps. In these profiles, the magnitude of the strong anomalies becomes evident. In Figure 3B, for example, the strong anomaly grouping for Zn-MS about 500 - 800 m from the mineralization stands out beautifully and is complemented somewhat by the other MS elements analyzed. THE CXT concentrations (Figure 5B) do not have the visual contrast as do those for the MS samples, although in total number of strong and moderate anomalies the CXT samples show a better multielement system. One could argue that by expanding the CXT scale (Figure 5B) and condensing the MS scale (Figure 3B) on the downstream concentration profiles, the visual contrast of the CXT samples would be accentuated and show up as well as the MS samples. This may be true, but as seen in Figure 6B, a comparison of Zn profiles for the three groups studied, the CXT scale is already expanded 200x over that of the MS scale and the definition of the MS strong anomalies is far better than that for the CXT strong

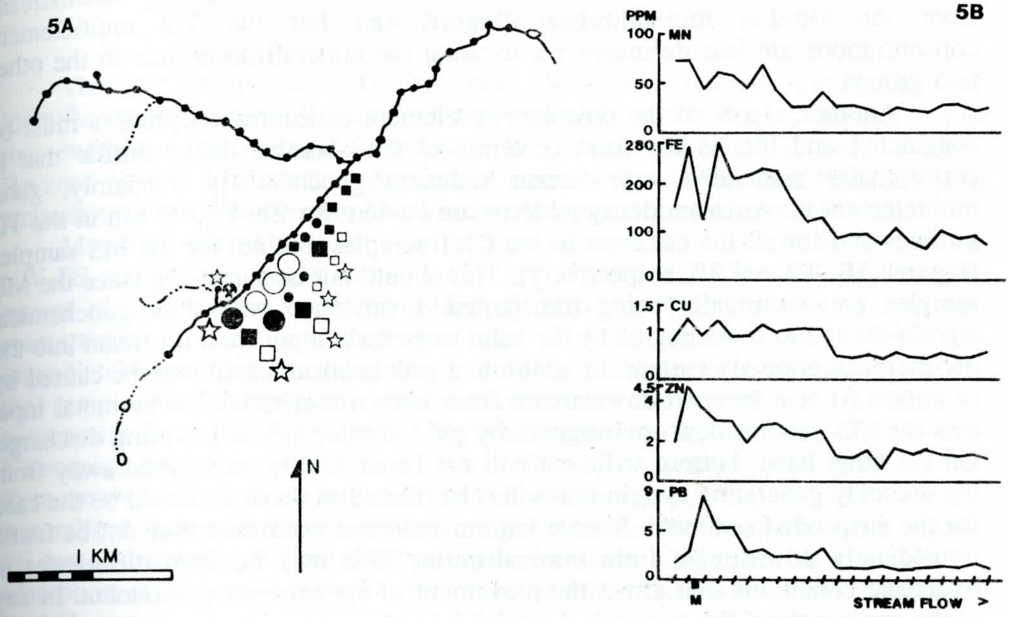


Figure 5. A. Geochemical anomaly map for the –100 mesh size fraction of bottom stream sediments, extracted with cold 3% HCl. Symbols as given in figure 3A. B. Downstream metal concentration profiles for the –100 mesh size fraction of bottom stream sediments, extracted with cold 3% HCl, and collected along the SW-NE flowing stream. M marks the position of the mineralization.

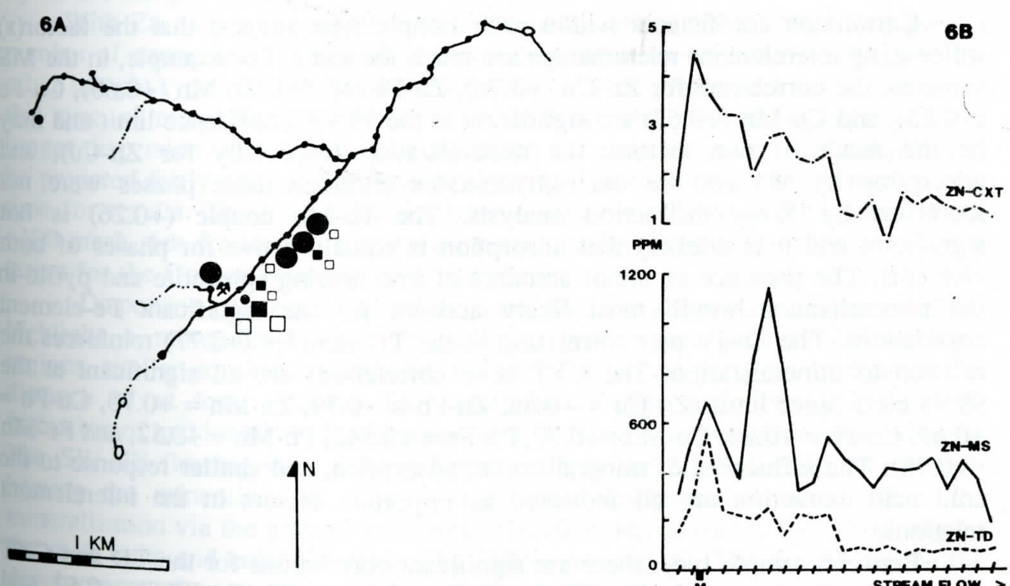


Figure 6. A. Geochemical anomaly map for Zn in the three sample types. In this figure, large symbols represent strong anomalies, small symbols represent moderate anomalies; full circles = Zn-MS; full squares = Zn-TD; open squares = Zn-CXT. B. Downstream concentration profiles for Zn in the three sample types collected along the SW-NE flowing stream. M marks the position of the mineralization.

anomalies. Zn-TD shows an important concentration immediately downstream from the Gordon mineralization (Figure 4B) but the TD multielement concentrations are less definitive for locating the mineralization than in the other two groups.

Another aspect of the downstream element concentration profiles must be considered and that is the form in terms of the classical decay profile that is conceptually associated with stream sediment geochemistry. Certainly, good multielement downstream decay patterns are evident for Zn, Fe and Mn in the TD samples and for all the elements in the CXT samples, but not for the MS samples (Figures 5B, 4B and 3B, respectively). This should not be surprising since the MS samples are continually being transported downstream with their geochemical signals diluted to some degree by the solid suspensate input from upstream into the geochemical anomaly system. In addition, a geochemical signal may be caused by or enhanced at a location downstream from known mineralization by metal input into the MS geochemical environment by groundwater inflow or spring discharge. On the other hand, bottom sediment will not be so readily transported away from the anomaly generating system nor will it be diluted as much as would be the case for the suspended sediment. Stream bottom sediment anomalies may not be found immediately downstream from mineralization. This may be from differences in hydraulic conditions that affect the movement of stream bottom sediment. In any study environment, the conceptual model related to stream sediment geochemical reconnaissance or follow-up work must be interpreted in light of the existing geological, geomorphological, and climatological conditions before selecting a zone for detailed work.

Correlation Coefficients

Correlation coefficients within each sample type suggest that the factor(s) influencing interelement relationships are much the same. For example, in the MS samples, the correlation for Zn-Cu (+0.76), Zn-Fe (+0.56), Zn-Mn (+0.60), Cu-Fe (+0.53), and Cu-Mn (+0.67) are significant at the 99.9% confidence limit and may be the result of two factors: the mineralization (especially for Zn-Cu), and adsorption by Mn and Fe oxides/hydroxides although these phases were not identified by X-ray diffraction analysis. The Fe-Mn couple (+0.26) is not significant and it is unlikely that adsorption is equally active for phases of both elements. The presence of minor amounts of iron-bearing sphalerite and pyrite in the mineralization would most likely account for the significant Fe-element correlations. The Zn-Fe pair correlation in the TD samples (+0.77) reinforces the relation to mineralization. The CXT suite correlations are all significant at the 99.95 confidence limit (Zn-Cu = +0.86, Zn-Pb = +0.79, Zn-Mn = +0.70, Cu-Pb = +0.69, Cu-Fe = +0.83, Cu-Mn =+0.70, Pb-Fe = +0.542, Pb-Mn = +0.52, and Fe-Mn =+0.78). The influences of mineralization, adsorption, and similar response to the cold acid extraction are all indicated as important factors in the interelement relations.

Between sample type, there are significant correlations for the TD and CXT samples for Zn, Fe and Mn (Zn-TD:Zn-CXT = +0.76, Fe-TD:Fe-CXT = +0.52, and Mn-TD:Mn-CXT = +0.90), and between Zn-MS:Zn-TD (r = +0.45) and Mn-MS:Mn-TD (r = +0.53).

X-ray Diffraction And pH

The clay minerals illite and kaolinite, plus quartz, feldspar and calcite were identified using X-ray diffraction of the suspended matter, but no mineral trends that suggest a mineralogical influence on element distribution were evident. Because Fe and Mn are not so strongly associated with the minerals found, an amorphous Fe-Mn phase may be present as suggested in the previous section.

Measured pH values were fairly constant at 7.1-7.3 for the waters of the NW-SE stream and ranged between 7.2 and 7.6 for the SW-NE stream waters except at one sample site where the pH was 8.0. The high pH sample site corresponds to the highest Zn-MS and Mn-MS values. The sample site immediately upstream from this high pH water had a stream water pH of 7.6 and the sample site immediately downstream had a water pH of 7.3. Mineral suspensates from these three sites have the highest Zn-MS contents (Figure 3B). No seepage or spring discharge was observed at or close to the high value site but there may be inflow from groundwater following the dip (30° SE) of the mineralized host rock. If inflow is taking place, it may be the cause of the higher pH and an increased Mn substrate precipitation which would provide a surface for adsorption of Zn associated with the inflow and abet the development of the high Zn-MS anomaly.

CONCLUSIONS

Beekmantown dolomite Zn and/or Pb deposits such as that represented by the Gordon property can be located for follow-up or detailed geochemical exploration by all the sample types evaluated in this study.

With Zn as a guide element to the mineralization, the Zn-MS has a greater number of strong anomalies with a greater concentration magnitude than Zn in the other sample types. However, the CXT suite may be preferred in exploration for this type of mineralization because it adds Pb to the anomaly system, yields a longer dispersion train (because of moderate Cu anomalies), and the samples can be prepared fairly quickly for AAS analysis so that there is a shorter turn around time for results. The least effective sample used was the total dissolution of the −100 mesh size fraction of the stream bottom sediment although it gives a good target for the Gordon property mineralization.

Dispersion patterns for all of the anomalies show that they will successfully highlight a mineralized zone in geologic, geomorphologic, and climatologic regions similar to that represented by the Gordon property to a downstream distance of at least 800 m. The structural geology undoubtedly affects the formation of and location of anomalies by its influence on groundwater movement. The SW-NE flowing stem follows approximately the N34°E strike of the rock hosting the mineralization. Any hydromorphic dispersion of metals from the mineralization via the groundwater would be expected to follow the 30° SE dip of the rock units and enter the stream by inflow or seepage. The results can be the development of or an increase of suspended particulate geochemical signal.

Single elements in all sample types studied will locate the mineralized zone, but the multielement anomaly dispersions are more effective in identifying such a zone.

The time and cost of geochemical analysis of the three sample types used in this study are equal. In terms of the time-cost perspective for sample collection and preparation for analysis, the cold acid extracted sample is preferred to the mineral suspensate sample which in turn is preferred to the total dissolution sample. However, as research on suspended sediment as a geochemical prospecting medium continues, the efficiency of sample collection will improve. When this improved efficiency is coupled with the mineral suspensate effectiveness in targeting mineralization, the suspensate may be the best sample in prospecting for specific types of ore deposits in given geologic-topographic-climatologic environments.

Because the mineral suspensate is a relatively untested sample type for geochemical prospecting, more research into its capabilities must be done. The good performance of the simple, low-cost, mechanical (hand-operated) vacuum-filtering system used in this study should stimulate such research in inaccessible as well as accessible areas. The research should include the response of mineral suspensates in different environments with similar mineralization, applicability of mineral suspensates to locate different kinds of mineralization with their attendant different geologies, the seasonal effect on element concentrations in different climatic regimes, and the contents of pathfinder elements that are not major elements in the ores sought (e.g., As, Sb, and Hg as pathfinders to hydrothermal Au deposits). The potential of the organic fraction of suspended matter in streams for geochemical exploration must be studied as well. Filipek and others (1981) reported that in drainage from the Magruder mine area, northern Georgia (USA) where Carpenter and others (1975) found pronounced anomalies for Cu-Zn-Pb on Mn- and Fe-oxide coatings of boulders, organic matter is an important sink for the ore metals in the boulder coatings. The order of concentrations was given as Cu, ORG > Fe-oxides > Mn-oxides; Zn, ORG = Mn-oxides > Fe-oxides; Pb, Fe-oxides > ORG > Mn-oxides. Finally, investigations into what is carried in specific components of a mineral suspensate such as metals adsorbed onto Fe and/or Mn oxide/hydroxide substrates should be carried out.

ACKNOWLEDGMENTS

Acknowledgments for field and laboratory assistance are extended to Richard Cox, Gabriela Siegel, and Edwin McFarland. Research into mineral suspensates as a guide to finding mineralization has been supported by The George Washington University Research Committee, The St. Joe Minerals Company, and The Anaconda Minerals Company (RIP).

REFERENCES CITED

Burman, J.O., 1983, Element transport in suspended and dissolved phases in the Kalix River: *in* Environmental Biogeochemistry (R. Hallberg, editor), Ecology Bulletin, v. 35, p. 99-113.

Carpenter, R.H., Pope, R.A. and Smith, R.L., 1975, Fe-Mn coatings in stream sediment geochemical survey: Journal of Geochemical Exploration, v. 4, p. 349-363.

Carserud, L., 1983, Dissolved and suspended heavy metals in stream water: *in* Environmental Biogeochemistry (R. Hallberg, editor), Ecology Bulletin, v. 35, p. 73-84.

Filipek, L.H., Chao, T.T. and Carpenter, R.H., 1981, Factors affecting the partitioning of Cu, Zn and Pb in boulder coatings and stream sediments in the

vicinity of a polymetallic sulfide deposit: Chemical Geology, v. 33, p. 45-64.
Herbert, P. Jr. and Young, S.R., 1956, Sulfide mineralization in the Shenandoah Valley of Virginia: Virginia Division of Geology, Charlottesville, Virginia, Bulletin 70, 58 p.
Medlin, J.H., Suhr, N.H. and Bodkin, J.B., 1969, Atomic absorption analysis of silicates employing LiBO2 fusion: Atomic Absorption Newsletter, v. 8, p. 25-29.
Perhac, R.M. and Whelan, C.J., 1972, A comparison of water, suspended solid and bottom sediment analyses for geochemical exploration in a northeast Tennessee zinc district: Journal of Geochemical Exploration, v. 1, p. 47-53.
Siegel, F.R., 1982, Suspended mineral matter as a stream sediment sample for geochemical prospecting (ext. Abs.): 9th International Geochemical Exploration Symposium, Abstracts Book, p. 101-103, Saskatoon, Canada.
Siegel, F.R., 1985, In-situ recovery of suspended sediments from streams: Journal of Geological Education, v. 33, p. 132-133.